TRANSACTIONS AMERICAN GEOPHYSICAL UNION

Volume 30 APRIL 1949 Number 2

SYMPOSIUM ON THE EARTH'S CRUST

(Dedicated to the memory of Alfred C. Lane)

Foreword

In planning the Twenty-Ninth Annual Meeting, held April 21-23, 1948, Merle A. Tuve was requested to arrange for a symposium on the Earth's crust. The following papers and abstracts present the essence of this symposium.

The last speaker was to have been Alfred C. Lane whose great career in science was ended by his death on April 15, 1948. At the meeting it was decided to dedicate the symposium to Dr. Lane's memory by reason of his many contributions to the Earth Sciences which will continue to serve as a stimulus for years to come, and by reason of the inspiration of his teaching to his many students who have risen to eminence in science.

RELATION OF MICROSEISMS TO METEOROLOGY

Marion H. Gilmore

Abstract--Microseismic data obtained by the Naval Aerological Service from its ten seismograph stations show that severe wind storms over water cause large sections of the Earth's crust to vibrate. These oscillations are called microseisms and have been recorded on sensitive seismographs up to 1600 mi from the source. The direction of propagation of certain dominant microseisms, especially those produced by hurricanes and typhoons, can be determined with great accuracy when they pass over a tripartite seismograph station. The microseismic method of storm detection as outlined in this paper is being developed by the Naval Aerological Service and is now used to supplement other methods of detecting and forecasting severe storms.

Microseisms are elastic waves that travel only along the surface of the solid Earth and, in that respect, they are similar to certain surface waves following earthquakes. They represent actual movement or displacement in the Earth's crust and can always be recorded on sensitive seismographs. There are many kinds of ground motions recorded on various types of vibration meters, such as those caused by earthquakes, by traffic, by natural or artificial explosions, by wind on trees, buildings, or high cliffs; but they differ from one another in type and period. The ground motion under discussion may be classed as either regular or irregular with periods ranging from two to six seconds. The amplitude of the waves occasionally increase to many times normal and continue large for a period of a few hours or many days. When they appear in a succession of groups of large buildups they are generally called microseismic storms.

The study made by J. Emilio Ramírez in 1938 and the more recent research carried on by the Naval Aerological Service, beginning in 1944, have proven that microseismic storms are directly related to certain types of meteorological phenomena. Microseismic data are now used by meteorologists to detect and track severe tropical storms and have been found very valuable in showing the intensity and position of the hurricane or typhoon, especially at night and when they are several hundred miles from land. The ideas submitted in this paper have been obtained directly from data collected from the Navy's ten seismograph stations in the Caribbean and Pacific during the last four years (see Fig. 1 and 2), wherein about 40 microseismic storms were recorded and studied in detail.

Before the source and cause of microseisms could be thoroughly investigated, it was first necessary to determine very accurately the direction from which they came. Several methods were carefully studied in order to compare their accuracy and dependability. One method was based on the assumption that they were composed entirely of Rayleigh type waves. If that were true it would be possible to get a rough estimate of the direction of propagation of microseisms by a comparison of the vertical with one of the two horizontal components. But all microseisms are not of the Rayleigh type wave. Other types have been found and now most observers agree

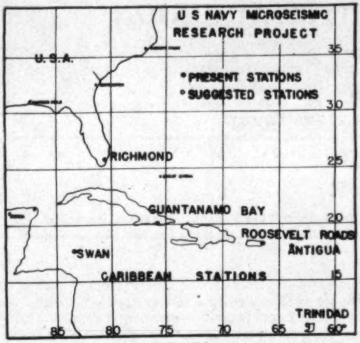
U S NAVY MICROSEISMIC
RESEARCH PROJECT
U.S.A.
• PRESENT STATIONS
• SUGGESTED STATIONS
RICHMOND
GUANTANAMO BAY
ROOSEVELT ROADS
ANTIGUA
SWAN
CARIBBEAN STATIONS
TRINIDAD
35
30
25
20
15
85
80
75
70
65
60°

Fig. 1--Location of Caribbean microseismic stations

with a recent statement by LEET [1947], "The type or types of waves represented by microseisms, however, is not yet fully understood." In actual practice it was found that the accuracy and reliability of this method fell far short of that desired and considered necessary by the armed forces for the finding and tracking of severe storms. Recent publications show that the accuracy obtained by comparing the vertical with one horizontal component has not increased and that the microseismic directions obtained by this method are never more precise than ± 22 1/2°.

Because of confusion surrounding the type or types of wave motions generally grouped as microseisms and the lack of precision in the first method investigated, it was decided that the Navy's Research Project would calculate bearings on microseismic storms by means of a tripartite station [GILMORE, 1946]. In this method, three seismographs, oriented in the same direction are used at the vertices of an equilateral triangle whose sides are about 2500 ft long. The direction of propagation of the wave front can always be determined by triangulation as the microseisms, regardless of their type or character, pass over the three instruments. Seven tripartite stations and three single instrument stations have been operated by the Naval Aerological Service in the development of the microseismic method of detecting and tracking of hurricanes and typhoons.

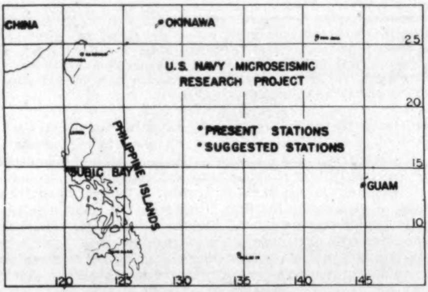
CHINA
OKINAWA
U.S. NAVY MICROSEISMIC
RESEARCH PROJECT
• PRESENT STATIONS
• SUGGESTED STATIONS
PHILIPPINE ISLANDS
SUBIC BAY
GUAM
25
20
15
10
120
125
130
135
140
145

Fig. 2--Location of Pacific microseismic stations

This work has been described in detail in recent papers [GILMORE, 1947; MURPHY, 1946; GUTENBERG, 1947; GILMORE and HUBERT, 1948] and the data presented leave no doubt of a direct relationship between microseismic storms and severe tropical disturbances. Case histories have been cited proving that hurricanes and typhoons, in some manner not yet fully understood, actually cause the floor of the ocean under them to vibrate. These data also prove (especially the 1947 results of typhoon tracking in the Pacific) that the energy of a storm is partly changed into strong vibrations in the Earth's crust and that these microseismic waves travel outward in more or less concentric circles, just as ripples do when a stone is dropped into a pool of water. When storms are concentrated over very small areas of strong winds, such as typhoons and hurricanes, the microseisms appear to come directly from that area of greatest turbulence, and that direction can now be determined with great accuracy. Present data indicate that storms do not cause microseisms when over land and that they are greatly reduced in amplitude when the storm passes over water that is not deeper than about 20 ft. On the other hand, microseisms show no change in amplitude or type when the water under the storm is at least 20 ft deep. This holds even when the depth exceeds 20,000 ft.

There are at least three important factors that limit the free transmission of microseismic waves produced by tropical storms. The first factor is the intensity of the storm, and the second is its distance from the recording station. It is evident that the energy in microseisms will be gradually lost as they pass through the resisting Earth's crust, just as ripples on a pond of water soon die out. That is why, unless a storm is extremely violent, the vibrations produced will seldom carry more than a thousand to 1500 mi, under the best of conditions. A third factor that limits the free transmission of microseisms can be termed, for lack of a better name, a microseismic barrier (see Fig. 3).

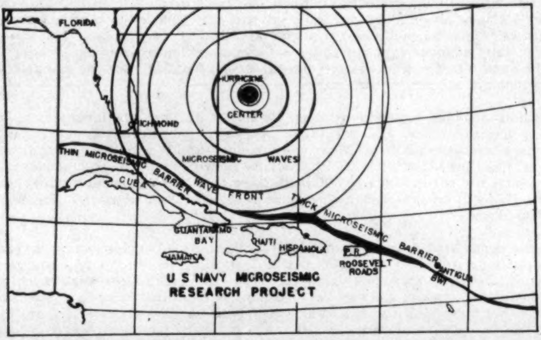

Fig. 3--Caribbean microseismic stations and the Barrier

Severe Pacific typhoons have been recorded by increased microseismic amplitudes for two years and accurately tracked by microseismic bearings when 1600 mi from a recording station. The seismograph, therefore, has a much greater effective tracking range on Pacific typhoons than on Caribbean hurricanes because certain zones have been found in the Caribbean that limit or altogether prevent the free passage of microseisms. Such zones, together with large geological blocks, have been mentioned in microseismic discussions for many years but, even so, the exact nature of this interference is not yet known. It might be a major fault system with vertical slippage on the order of 40,000 to 50,000 ft where one side is unconsolidated marine sediments, and the other side, basement rock. Such faulting could possibly absorb or reflect a major portion of the energy of microseisms. On the other hand, Byerly reports that his teleseismic station at Berkeley is located between two major rifts and that few stations are more troubled with microseisms than his. In the Pacific area around Guam, Okinawa, and Manila, these barriers are almost completely lacking and the vibrations gradually die out as their distance from the recording station increases upward to 1600 mi.

At present, it is impossible to do more than guess concerning the exact nature of the microseismic barrier. In the Navy's research work, it is now sufficient to know that such zones exist and to know their location, in order that proper allowances be made for them in the final interpretation of microseismic storms. A very definite barrier (see Fig. 3) almost completely absorbs or reflects the microseisms caused by severe hurricanes passing north of the Navy's tripartite station at Roosevelt Roads, Puerto Rico. Little is known of its eastward extension except that it probably passes south of Antigua Island and north of Trinidad. It extends westward and passes into the Gulf of Mexico somewhere between Key West and Havana. Storms passing north of Cuba lose approximately one-half or more of the microseismic amplitude in crossing the boundary before reaching the station at Guantanamo Bay, Cuba. The extension into the Gulf of Mexico is not well located. Microseisms caused by hurricanes along the south Texas coast will not reach the Florida station near Miami. On the other hand, they are well recorded for all storms passing over south Florida and as far as the New Orleans area. This and other microseismic barrier zones will gradually be located and accurately mapped as more data are collected and studied.

There are also other factors that must be considered before it is possible to operate a tripartite station at its highest efficiency in locating and tracking severe tropical disturbances. The locating of storms by means of microseismic bearings is one of the most important functions of a tripartite station. The shape of the station triangle or type of microseisms recorded are not important. On the other hand, the similarity of wave motion recorded on the three seismographs of a tripartite station is of vast importance in the final determination of the direction of propagation of the microseisms. Extreme care must be taken with the entire instrumental setup to insure that each microseismic wave is accurately recorded on all three seismographs of the tripartite station. Before a bearing can be calculated it is necessary to identify each individual wave on the three records in order to measure the small differences in times of arrival. The 1944 station was a right isoceles triangle with each leg about 18,000 ft long. Many of the microseismic waves would not carry that distance without distortion in form that made them difficult to identify and measure. Since then, the distance between stations has been gradually reduced to approximately 2500 ft and, to simplify the calculations, equilateral triangles are used where possible. The three instruments of a set are carefully adjusted to the same period and magnification so that, with the short distance between stations and a fast recording drum speed, almost identical records are obtained on the three seismographs of each tripartite station.

It is therefore possible to secure bearings on nearly all types of microseismic storms and to determine very definitely where they come from, especially when cross-bearings can be obtained from two or more tripartite stations. When due consideration is given to microseismic barriers and areas of very shallow water, the amplitude of the microseisms can be considered a direct function of the intensity of the winds and their distance from the recording station. An increase of wind over the ocean will always cause an increase of microseisms at nearby seismographs in the same geological area.

An increase in the wind over the Florida Keys will cause a corresponding increase in the amplitude of microseisms at the south Florida station. Microseisms of this type are very common at all stations and are always irregular in form. They often reach large amplitudes, depending entirely on the force of the winds, and they never travel far from their source. It would be impossible for them to set up oscillations in a whole continent. Research in recent years has clearly demonstrated that the only accurate and positive way to determine the direction of propagation of microseisms is by measurements made as they pass over a tripartite station. Therefore, it seems unwise even to suggest that they come from any definite direction or from any central source unless that direction has been obtained by the tripartite method. Microseisms of this type have been recorded on stations in south Florida, eastern Cuba, and eastern Puerto Rico, and the bearings calculated at each station never point to a central source. The microseisms appear to come from various points of high wind very near the recording station. Their irregular type make it difficult to get good bearings and this irregularity is likely caused by the microseisms coming from first one direction and then another, making a very confused, criss-cross pattern. It is definite that they do not come from any pinpoint source. A severe typhoon, hurricane, or area of high winds from a frontal system, will produce the same type microseisms when the distrubance is over or very near the recording station because the actual source is around the station rather than a specific direction from it. The amplitude of such microseisms sometimes reach 100 times normal and are of little value to a forecaster of severe storms when it is over his station.

On the other hand, various forecasting agencies of the Army, the Air Force, the Navy, and the Weather Bureau are extremely interested in the type of microseism that can describe a severe tropical or extra-tropical storm when it is 100 to 1600 mi from the recording station. Microseismic storms of this type are regular in form and have one or more good buildups per minute. The detection and tracking of typhoons in 1947 leave no doubt concerning the value of microseismic data. Microseisms caused by Pacific storms and Caribbean hurricanes contain much valuable information concerning the intensity of the storm. Tripartite stations, such as Guam, furnish excellent bearings that point directly into the storm area. The actual position of a storm can be accurately found by cross-bearings from two or more tripartite stations that record it. Recent data from the new station at Subic Bay, near Manila, show that a storm from the other side of the Philippine Islands can be accurately tracked as it moves along its course. This fact indicates that there are no barriers in the Philippine Island group east of Manila that tend, in any manner, to impede the free passage of microseisms.

Operational techniques have been developed and improved to a point where several additional tripartite microseismic stations in the Pacific typhoon belt (see Fig. 2), would greatly improve existing methods of detecting and tracking severe storms. Such a system of stations would undoubtedly reduce the overall cost of forecasting typhoons and, at the same time, keep an ever constant 24-hour watch over much larger areas of the oceans than ever before. This is especially true in the Pacific because of the large distances microseisms travel from typhoon centers. It is also true for the Caribbean stations when they are located on the same geological block as the hurricane, or rather on the same side of the microseismic barrier (see Fig. 1 and 3). Therefore, before this method of detecting hurricanes can be fully realized in the Caribbean it will be necessary to relocate some of the existing stations and to add several additional tripartite stations, as suggested in Figure 3.

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(Manuscript received May 17, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 21, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

ABSTRACT OF DISCUSSION

(Balch Graduate School of the Geological Sciences, Contribution No. 453)

B. Gutenberg (California Institute of Technology, Pasadena, California)--As early as 1921 the writer has pointed out that the propagation of microseisms in Europe is clearly affected by boundaries of major geological units. For example, microseisms originating near Scandinavia are propagated far into Asia without much loss of energy, but their amplitudes decrease considerably in southerly directions where surface layers of different geologic age and with different elastic constants are passed by the waves. Similarly, microseisms are well propagated across large parts of the United States. On the other hand, records of microseisms produced by hurricanes in the Caribbean area and recorded there at the stations of the United States Navy Department show that the amplitudes of the microseisms decrease rapidly from one island to another. In the Pacific the station at Guam of the Microseismic Project of the United States Navy records microseisms from typhoons with their center far away in the area east of Japan. Japanese scientists have pointed out that typhoons in that area are accompanied by large microseisms at the stations along the east coast of Japan but only by insignificant microseisms on the west coast of Japan. There is some indication that typhoons moving westward south of Guam produce relatively irregular and small microseisms at Guam until they cross the Andesite Line near the Caroline Islands. Thus far, in all instances where the microseisms are propagated over long distances without much loss of energy, stations and cause are on the same geological unit. On the other hand, where amplitudes of the microseisms decrease rapidly with distance there is some evidence that they have crossed one or more discontinuities separating areas with different seismic velocities. Faults need not reduce appreciably the amplitudes of microseisms as long as velocities on both sides of the fault zone are approximately the same, and as long as the fault zone is narrow as compared with the wave length. Study of the decrease of amplitudes of microseisms based on more extended data can be expected to give more information on boundaries of distinct units of the Earth's crust.

PRE-HISTORY OF THE EARTH'S MAGNETIC FIELD

E. A. Johnson

Abstract

In order to determine the origin and nature of the Earth's magnetic field and to test the various hypotheses which have been advanced to explain the field, it is desirable to determine the history of this field throughout geologic time and to investigate more carefully its spatial variations, both inside and outside the Earth's surface. The research described was concerned with the determination of the history of the Earth's field as it can be reduced from the present polarization of crustal material. Unconsolidated fresh- and salt-water sediments have been investigated. These sediments are in the form of clays and offer one of the simplest types of polarization since the clays can be redeposited under laboratory conditions. A particularly lengthy investigation of the polarization of glacial clays has been made, together with measurements on core samples of sediments from the Pacific. From a study of anomalous deposits in the glacial clays, the geologic stability of the polarization of these clays has been established over geologic time. From the measurements of the glacial clays, it is concluded that the Earth's field has not changed substantially in direction or intensity during the last 15,000 years. From measurements of the Pacific cores, it is tentatively concluded that the direction and intensity of the Earth's magnetic field has probably remained substantially constant during the last million years. These results are consistent with the fundamental theory proposed by Schuster, Babcock, and Blackett, but do not provide evidence to support this theory.

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(Abstract received November 16, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 21, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust.)

STRUCTURE OF THE MID-ATLANTIC RIDGE

Maurice Ewing

Abstract

Atlantis Cruise 150, sponsored by the National Geographic Society, Columbia University, and the Woods Hole Oceanographic Institution, was a two month trip to Bermuda, the Azores, and the Mid-Atlantic Ridge in the summer of 1947. A continuous record of depth beneath the ship along the track gave significant topographic information and provided a valuable guide to the choice of locations for measurements of other types.

Thirty-five sediment cores were taken with the Stetson-Hvorslev apparatus, about a ton of rocks were dredged up, and several excellent seismic refraction records were obtained at various stations. Seismic reflection measurements were made at intervals of about ten miles along most of the track and some work was done on photography of the ocean floor. Hydrographic and biological observations were made, including a number of hauls with the Blake trawl.

From the Azores to about Lat 29°N the central highland of the Mid-Atlantic Ridge is about 60 to 200 mi wide. Except for a few deep linear trenches, this part of the Ridge lies entirely above 1600 fathoms and is rugged. Sedimentary cover of this part of the Ridge is very thin. On both sides of this central highland there is a broad system of terrace-like feature extending from 1600 fathoms to 2500 fathoms. The reflection seismograph shows the depth beneath the ocean floor to the principal buried reflector on the flanks and terraces of the Ridge to be as great as 3000 ft, but only 100 ft on the central highland or in the main western basin. To the west of the terraces is a gradual descent into the main basin of the western Atlantic which lies at 2900 fathoms. The rocks which were dredged up from the central highland of the Ridge have been examined by S. J. Shand who states that the "collection confirms the predominance of basalt and makes at least two notable additions to the record, namely an abundance of serpentine and a mylonitized anorthositic gabbro."

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(Abstract received December 10, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 21, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust.)

SEISMIC STUDIES OF GEOLOGIC STRUCTURE OF THE OCEAN FLOOR

J. B. Hersey and Frank Press

Abstract

From April 1946 to April 1948, significant advances were made in the study of the structure of the ocean basins using the seismic method. Swedish scientists have reported results obtained with a reflection method employing a piezoelectric hydrophone as the receiver and small explosive charges as the source of seismic energy. They report having recorded reflections from surfaces below the ocean floor. They demonstrate penetration of oceanic sediments by interpretation of several records of shots detonated at the same place but at radically different depths in the water. Their recording system has essentially a flat frequency response.

A group at the Woods Hole Oceanographic Institution developed a reflection technique in the fall of 1946 employing small explosive charges and a piezoelectric hydrophone. In this method, penetration of bottom sediments is demonstrated by analysis of the frequency composition of the reflected energy from a single shot. Energy reflected at the ocean floor is reflected over a wide range of frequencies, while reflections from horizons below the ocean floor are recorded only at frequencies well below 1000 cps. Over 100 sets of records were made on which reflections below the ocean floor were recorded.

In the summer of 1947 a group from the Geology Department of Columbia University further developed this method so that observations could be made from a ship underway.

Since that time these latter two groups working in cooperation have made several improvements in recording and shooting techniques and in the process have taken reflection records at frequent intervals on several cruises over many hundreds of miles in the northwestern Atlantic and on one cruise from the Strait of Gibraltar to Woods Hole. The work of accumulating additional data in the north Atlantic is continuing as opportunity affords. To date approximately 1500 reflection stations have been occupied by both groups. Analysis of the results is underway and will be prepared for publication. A paper describing the results of the early work and suggesting a system of classification of bottom reflection records was published in the Transactions of the American Geophysical Union, February 1949, pages 5-14.

Somewhat similar methods have been used successfully by the Columbia University group in refraction seismic studies in both shallow and deep water. Considerable data have been accumulated and the results are being prepared for publication.

Analysis of reflection records to date indicate that a principal reflecting horizon is to be found whose reflected arrival lags the bottom reflected arrival by times from less than 0.1 sec to more than 1.5 sec. What this represents in thickness of material to this horizon will not be known until the velocity of seismic waves in the bottom material has been determined. Any reasonable estimate, however, indicates negligible cover in some parts of the deep ocean and thickness of the order of several thousands of feet in other parts.

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(Abstract received November 9, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 21, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust.)

SIGNIFICANCE OF RADIOACTIVITY IN GEOPHYSICS - THERMAL HISTORY OF THE EARTH

Wm. D. Urry

Abstract--It appears that the effects on the Earth's thermal history of the exponential decay of the sources of atomic (radioactive) heat within the Earth are such that the upper crust of the Earth was heating in its early history and that subsequent cooling has been more nearly linear than had been supposed. In the deeper parts of the crust and below, the thermal history has been complex with simultaneous heating at one depth and cooling at another depth. Temperatures in the past beneath a Pacific-type ocean have not varied in the same manner.

Recent determinations of the radioactivity of ultrabasic rocks and iron meteorites are such that a monomineralic olivine mantle and an iron core will have remained essentially at the initial temperatures.

The Earth, now presumably differentiated, may have had an earlier existence as an undifferentiated body formed by the coalescence of particles. Its rebirth as a molten, differentiated, and later frozen body is possible if the total radioactive content of the Earth was previously distributed uniformly and if its dust-cloud origin dates back about seven billion years. A parental planet from which might have come the asteroids and meteorites may have disrupted because of stresses accumulated by internal heating.

Foreword--This paper is essentially an abridgement of a dissertation on the Thermal History of the Earth by Wm. D. Urry and G. Comenetz. That paper, which has been circulated privately, treats in detail the mathematical problem of computing the surface heat flow as a function of time, and, to a lesser extent, the temperatures as function of depth and time, for three different assumed Earth models; and it presents an exhaustive and critical survey of the data required for this and similar problems. This abridgement gives the principal experimental data and assumptions that enter into the computations and give a summary of the results, and shows the more important ways in which these data and assumptions and the consequent results differ from those of previous investigators. Certain implications of the results are then pointed out and finally the result of applying the same methods to two other problems is briefly sketched.

Introduction--The thermal state of the Earth has been considered in more or less detail by a great many writers during the past hundred years, although few have considered the thermal history, since they were interested mainly in the initial, present, and final equilibrium thermal state of our sphere.

With the discovery of a widespread distribution of radioactive matter in the surface rocks of the Earth, a very important source of heat within the Earth was first recognized. It was soon found that on any reasonable assumption regarding the distribution of such matter in depth, the major part of the heat, being lost from the surface of the Earth, must come from radioactive disintegration. This atomic heat accounts for from 3/4 to 4/5 of the present total surface heat flow. HOLMES [1915], ADAMS [1924], JEFFREYS [1929], and GUTENBERG [1939], each making somewhat different basic assumptions and recognizing the sources of atomic heat within the Earth, have published curves showing present temperatures against depth for some hundreds of kilometers below the surfaces of their particular Earth models. At great depth within the Earth, actual temperatures cannot be determined by any presently available methods; only the rise in temperature during the life of the Earth can be calculated from the supposed concentrations of the radioelements at these depths; that is, actual temperatures at great depths today can be estimated only after the assignment of the initial temperatures at such depths.

All previous investigators have, for simplicity, assumed a constant rate of generation of atomic heat; that is, they assumed that the source of atomic heat was inexhaustible. Actually the amount of any radioelement decreases exponentially with time, and the rate of generation of atomic heat arising from that radioelement decreases correspondingly. Thus the total rate of generation of heat from the uranium-238 series, the uranium-235 series, the thorium series, and from potassium-40 has decreased by roughly 30 per cent in 2000 million years. HOLMES' [1946] recent calculations indicate an age for the Earth of about 3000 million years. In this period the decrease in heat generation has been nearly 50 per cent. These figures are quite independent of the concentrations of the radioelements and are dependent only on the relative abundances of the radioelements and their established nuclear properties. It is of interest to investigate what effect this decrease in heat from this source will have on the thermal history of the Earth and how the picture of the true thermal history will differ from that based on the erroneous assumption of an inexhaustible supply of radioactive heat within the Earth. This is the task that was undertaken in the dissertation by Urry and Comenetz mentioned above, a synopsis of which forms the main part of the subject matter of this paper.

In a sphere, the flow of heat by conduction past a point at a distance r from the center, and having a momentary temperature, θ, is given by

$$\rho c\, d\theta/dt = (1/r^2)\partial\ (kr^2\ d\theta/dr)/\partial r + H\ (r, t) \quad \ldots\ldots\ldots\ldots\ldots (1)$$

where ρc is the volumetric heat capacity and k the thermal conductivity. H is the rate of generation of atomic heat and is a function of r, and of the time t, because the source of atomic heat within the Earth is not inexhaustible but diminishes exponentially with time.

The most recent and comprehensive application of (1) to the problem of the thermal state of the Earth is that offered by SLICHTER [1941]. Slichter, however, considered H as a function of r only and neglected the decrease in H with time t. Thus Slichter's values for the surface-heat flow in the final thermal equilibrium state of the Earth are finite and final interior temperatures are very high for most Earth models. It is shown below that this cannot be true although the thermal state of the Earth has tended towards these conditions in the past.

The unpublished expressions for the temperature θ, at any given depth and time, and the surface-heat flow F_a at any given time, derived from (1) for a layered Earth by my colleague, G. Comenetz, are extremely complex and laborious to handle by manual computing machines. The expression for the heat flow F_a has to be solved for only a single depth, the surface, and so far it has proved feasible to study only the variation of the surface-heat flow with time. There is much to be gained from a knowledge of the history of the heat flow because in the upper portion of the crust where the thermal gradient is constant, the temperature will vary in the same manner as the heat flow.

Solution of the equation--In order to solve numerically the integrated form of (1) for the contribution to the surface-heat flow F_a (called the partial heat flow hereafter), arising from the disintegration of a given radioelement (or its series) it is necessary to know (1) the thermal diffusivity, (2) the depth of the discontinuities between the shells, and (3) the concentration of the radioelements in each shell. In addition it is necessary to know the nuclear constants such as the decay constant and the energy released in all of the disintegrations for each of the three radioactive series and in each shell. In addition it is necessary to know the nuclear constants such as the decay constant and the energy released in all of the disintegrations for each of the three radioactive series and potassium-40.

In any known integrated form of (1) it is necessary to consider the thermal diffusivity, $k/\rho c$ as a constant. From a critical study in the original dissertation it appears that while the conductivity k increases with depth, the volumetric heat capacity increases to about the same percentage extent and the thermal diffusivity is remarkably constant at a value close to 0.007 cm^2/sec^{-1} at least to depths of the order of 600 km.

The factors (2) and (3) above are prescribed by the particular Earth model under study and insofar as necessary for the argument will be evaluated below. In order to avoid undue bias in the selection of the data of radioactivity, two types, (a) and (b), of the distribution of radioactivity have been studied. Type (a) assigns the maximum concentrations that can be considered from a critical survey of the modern data; type (b) assigns the minimum concentrations. The most probable values appear to be closer to the type (b) distribution.

The only assumptions made are that the surface temperature has remained always constant at 0°C (the exact value is immaterial) and that loss of heat from the interior to the surface is by conduction only. Therefore the calculated surface-heat flow is entirely independent of any measured value of the heat flow or thermal gradient.

The equation must be solved for each value of the decay constant for each value of the time t and each value of the thickness of a surface shell. To the sum of the partial heat flows, determined for a complex model of the Earth as indicated below, must be added the heat flow arising from the Earth's original heat. This can be calculated from the equations given by FOURIER [1878] but it is difficult to determine because of the assumptions that must be made concerning the initial distribution of temperature within the Earth. The problem is simplified, however, by the fact that initial temperatures in the crust are of major importance for the first tens of millions of years only and that essentially no heat has been lost to this day from below about 700 km.

Partial heat flows--Calculations for the partial surface heat flows F_a, defined above, in terms of H_o which is the rate of generation of atomic heat at t = 0, are shown in Figure 1 as a function of time for a shell 101.4 km thick at the surface of the Earth containing material with a uniform concentration of radioelements and no source of atomic heat below. F_a/H_o must be considered separately for each radioelement except when the supply is considered inexhaustible in which case the total value of F_a/H_o is given by the broken curve. It is apparent that the tendency for each partial heat flow to increase to the steady-state equilibrium value (dashed curve) that it would attain if the source of atomic heat were inexhaustible, is overcome by the decrease in the rate of production of atomic heat. This results in maxima in the partial heat flows, such maxima occurring earlier in the history of the Earth the shorter the half-life of the controlling radioelement. Calculations similar to those for Figure 1, but made for deeper shells and longer periods, show that the Earth will have little or no internal source of atomic heat and will have lost practically all of its original heat after 150 billion years. In the final and equilibrium state the surface-heat flow will be zero and the temperature will be everywhere the same as that at the surface.

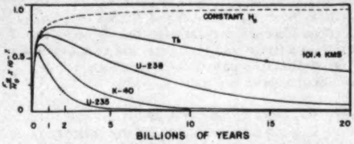

Fig. 1--The partial heat flows F_a, in terms of H_0, the rate of generation of atomic heat at $t = 0$, as a function of time

By constructing a number of curves similar to Figure 1 for shells of different thicknesses, each extending from a given depth to the surface, it is possible to interpolate and compute the surface heat flow for a complex model of the Earth consisting of any number of layers each containing a given concentration of the radioelements.

Distribution of heat sources--In order to compute the surface-heat flow it is necessary to know the present concentrations of the radioelements within the Earth from which values of H_0, the rate of generation of atomic heat at $t = 0$ can be calculated. This requires a model of the interior structure of the Earth and a knowledge of the content of the radioelements associated with the various types of material within the Earth.

Earth models--Three Earth models have been studied. ADAMS and WILLIAMSON [1925] proposed a structure of the interior of the Earth, the essential feature of which, for the present argument, consists of a silicate mantle of monomineralic magnesium-iron silicate. WASHINGTON [1925] based his arguments for an Earth model almost entirely upon analogy with the various kinds of stony and iron meteorites and thus proposed a silicate mantle composed of the meteoritic silicates. Likewise DALY [1943] draws a parallel between the composition of the achondritic and chondritic stones and suggests a mantle of meteoritic silicate. Other differences between these three Earth models have to do chiefly with the depths of the discontinuites. The arguments of Adams and Williamson were entirely independent of any consideration of the radioactivity associated with the mantle, core, or crust of the Earth and were based largely on the measured and observed velocities of seismic propagation. It is therefore interesting to find that in the light of recent work by DAVIS [1947] on the radium content of the dunites, the negligible radioactivity associated with unaltered olivine generates no embarrassing amount of atomic heat at great depth; whereas the radioacitvity associated with the stony meteorites would cause seemingly impossible rises of temperature.

Surface heat flow-- Figure 2 shows for Adams' Earth model the variation of the total surface-heat flow with time derived from data presented in Table 1. Figure 2 serves to illustrate the continuing decrease of the heat flow in the future and the error of assuming a constant rate of generation of heat. Figure 2 also demonstrates the peculiar feature of a rise to a peak value in the heat flow in the early history of the Earth, more easily seen in Figure 3, which shows that this feature is common to all three models and to the high (a) and low (b) types of radioactive distribution. (From a survey of the available data, the lowest assignment of radioactivity that can be made for the meteoritic silicates is as follows: anchondritic meteorites, $Ra = 0.07 \times 10^{-12}$ gm/gm, $Th/U = 3$, $K/U = 1 \times 10^4$; chondritic meteorites, $Ra = 0.05 \times 10^{-12}$ gm/gm, $Th/U = 3$, $K/U = 1 \times 10^4$. These values should be compared with the values for dunite in Table 1.)

The past thermal history of the crust can be divided into four periods. The surface-heat flow increased for the first 300 to 500 million years; temperatures in the upper part of the crust were increasing. Following this period, the flow of heat remained more or less constant for about 100 million years; temperatures in the upper crust were high and constant. For the next 1000 million years, to the close of the Pre-Cambrian, the heat flow decreased linearly; the crust was cooling linearly. At the close of the Pre-Cambrian there was a rather sudden slowing-up of the decrease in heat flow which has been maintained to the present.

Comparison with measurements--The average surface-heat flow from the few reliable measurements by BULLARD [1939], BENFIELD [1939 and 1947], and BIRCH [1947] is 1.32×10^{-6} cal $cm^{-2}sec^{-1}$. The calculated present-day value for Adams' Earth model with (b) type radioactive distribution is 1.38×10^{-6} cal $cm^{-2}sec^{-1}$. One might suppose that this agreement of two quite independently determined values offers strong evidence in favor of accepting Adams' model rather than the meteorite models of Washington and Daly. However, this is not the case because the calculated surface-heat flow today is but little dependent upon the amount and distribution of the radioactivity below 700 km. It is chiefly dependent upon the total radioactive content of the upper 40 to 100 km but not to any great extent upon the manner in which this total is distributed in this upper layer. The most that can be argued is that Adams' Earth model is not inconsistent with the measured values of the surface-heat flow.

Other ages for the Earth--Figure 4 presents the calculations for three different ages since the solidification of the Earth. The broad features discussed above are independent of the Earth's age but the important dates in the thermal history occur at different times on the geologist's time scale.

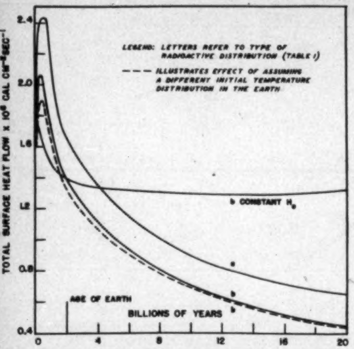

Fig. 2--Variation of the total surface heat flow with time for Adams' Earth model

Thus for an Earth that is 3000 million years old, according to Holmes[d], the surface-heat flow, and consequently the temperatures in the upper crust, have decreased almost linearly since early Pre-Cambrian times.

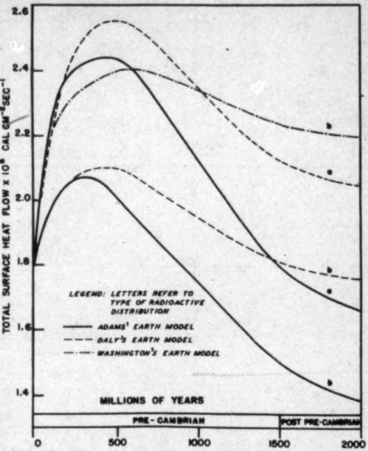

Fig. 3--The total surface heat flow in the past

Table 1--The radioactive data for Adams' model of the Earth

Depth of discontinuity	Shell and composition	Type	Density	Ra $\times 10^{12}$	Th/U	K/U $\times 10^4$
km			g/cm³	gm/gm		
0-15	Granite-	a	2.8	1.05	5	1.2
	Granodiorite	b	2.8	0.84	3.2	1.1
15-60	Changing from acidic					
	to basic 15-30 km	a and b	2.9	0.49	3.0	0.9
	30-45 km	a	3.1	0.34	4.0	1.1
		b	3.1	0.31	3.5	1.2
	45-60 km	a and b	3.3	0.20	3.4	0.9
60-1600	Dunite	a and b	3.9	0.0044	3	1
1600-3000	Pallasite[c]	a and b	7	0.013	3	0.5
3000-6370	Nickel-iron	a	10	0.023	3	0
		b	10	0.003	3	0

[c] More recent studies indicate a much thinner pallasitic layer between about 2700 and 2900 km, thus extending the silicate mantle down to 2700 km and the core up to 2900 km. Recent measurements of the radium content of pallasites indicate a value nearer to 0.003 and measurements of siderites corroborate the lower value of 0.003, (b) type, for the nickel-iron core. Neither of these revisions will have any appreciable influence on the surface-heat flow for many billions of years.

Moreover the temperatures in the crust in the early history of the Earth were correspondingly higher than in the case of the 2000 million-year age although the peak was passed in Pre-geologic time rather than in early Pre-Cambrian time. The term Pre-geologic is employed to connote the time prior to the crystallization of the oldest dated minerals and it is not implied that rocks and sediments of this era are nonexistent today. However, from what follows below it would appear that the dynamic movements and changes at the Earth's surface were greatly accelerated over present rates in this early era.

Contraction of the crust--The rate of cooling of the crust has been more nearly linear than had been supposed previously. This removes one of the chief objections raised by Holmes [JEFFREYS, 1929] to the theory of mountain building by contraction of the crust, namely, that the periodicity of orogenic movements has remained constant whereas it should have changed if the cooling rate were not linear. However, the previously estimated amount of contraction was considered by many to be barely sufficient to account for the observed crustal shortening. The present considerations must

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[d] Holmes' result was published after many of these computations had been made, hence much of the discussion here centers on an age of 2000 million years.

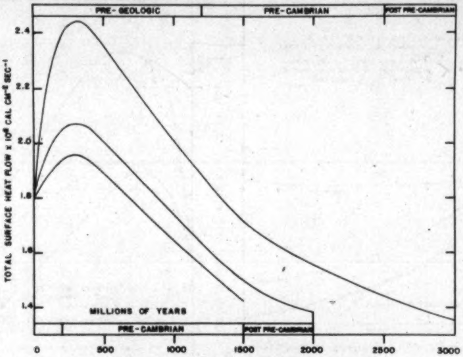

Fig. 4--The total surface-heat flow in the past for three ages of the Earth

affect the previous estimate of the amount of contraction but by how much cannot yet be stated quantitatively because depths are involved much below those for which any near-surface temperatures can be extrapolated.

Past crustal temperatures--Changes at the Earth's surface should be greatly accelerated when the rocks at a given level attain a certain critical temperature. In an anhydrous granitic crust this temperature could not be much below 800°C but the presence of water undoubtedly lowers this limit by an unknown amount. Table 2 gives the depths of the isotherms for three temperatures in the past. Even if one does not accept an appreciable effect of water in lowering the temperature at which occur the phenomena associated with approach to anhydrous melting, the 800°C isotherm was close enough to the surface compared with the present depth to warrant the assumption of a great rapidity of events at the surface in Pre-geologic and possibly early Pre-Cambrian times. If a temperature of 600°C is significant, the early crust must have been in an almost continuous state of turbulence.

Table 2--The past and present depths of three isotherms beneath a continental crust in kilometers

Time from present	Isotherm			Geologic time scale
	450°C depth	650°C depth	800°C depth	
million yr	km	km	km	
present	15	21.5	26.5	Post Pre-Cambrian
500	14	20	25	Post Pre-Cambrian
1000	13	19	23	Pre-Cambrian
1500	12	17	21	Pre-Cambrian
1800	10.5	15.5	19	Pre-Cambrian
2000	10	14.5	18	Pre-geologic
2500	8.5	12.5	15.5	Pre-geologic
2700	8	12	15	Pre-geologic
2900	9	13	16	Pre-geologic
3000	Age of the Earth			

Flow of heat beneath the ocean--Previous discussion has been confined to a continental crust. The history of the flow of heat beneath a Pacific-type ocean, where presumably a granitic shell is lacking, has differed appreciably from that out of the continental masses, as shown in Figure 5. Interpreting Figure 5 in terms of upper crustal temperatures, temperatures beneath the ocean have always decreased; beneath the continents they increased by about 15 per cent in the first 300 million years. In the same period the temperatures below the ocean fell by about the same

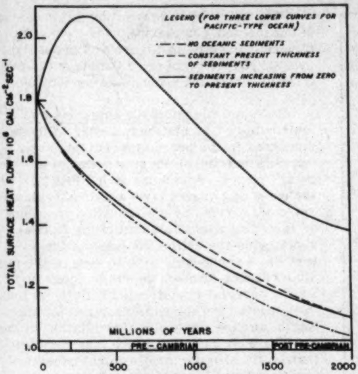

Fig. 5--Comparison of the total surface-heat flow in the past beneath a Pacific-type ocean (lower three curves) with that in a continental crust (upper curve)

amount. In Pre-Cambrian time, upper crustal temperatures under the continents fell nearly linearly and at a greater rate than beneath the ocean, where, in addition, cooling was far from linear. These differences are probably significant in the problem of the origin and history of the continents and ocean basins and must certainly have some bearing on the circumferential tectonic features of the Pacific ocean. The case for an age of 3000 million years is similar except that temperatures beneath the ocean remanied nearly constant for the first 300 million years while they rose by 36 per cent in the continental crust.

Past temperature changes at depth--All analyses of the problem of the Earth's interior temperatures agree in attributing to the Earth an enormous thermal inertia with all indication that little or no heat has escaped from below 700 km even in 3000 million years. One can therefore compute in a simple manner the rise above the initial temperature caused by the generation of atomic heat. The initial temperature is, according to **ADAMS** [1924], the solidification temperature of the material existing at a given depth under the pressure at that depth. At 1600 km the atomic heat generated in 2000 million years in the meteorite models of Washington and Daly would have caused a rise of between 1200° and 1800°C depending upon the necessity to supply the latent heat of fusion. For various reasons this rise appears to be unacceptable. (Two such reasons appear to make such a rise in temperature above the solidification temperature untenable are the strength and rigidity demanded by the triaxiality of the Earth's figure and the passage of the distortional (s) seismic wave at this depth. The central core appears to be a fluid and although under considerably greater pressure it apparently fails to transmit the distortional wave.) On the other hand, the heat generated by the radio-elements associated with olivine would be insufficient to provide more than about one third of the heat of fusion or to cause a rise of more than 150°C if solidification had ceased just short of crystallization.

In the supposed pallasitic layer at 2700 km the accumulated atomic heat would be insufficient to provide more than a fraction of the heat of fusion or to cause a rise of more than 100°C if the material had never lost its heat of fusion.

Recent measurements of the radium and thorium content of the iron meteorites by **ARROL, JACOBI, and PANETH** [1942] and by G. L. Davis (unpublished) at the Geophysical Laboratory indicate an even lower rate of generation of atomic heat in an iron core than in the silicate (olivine) mantle.

Thus temperatures deep in an Earth, differentiated to the extent proposed by Adams, have remained essentially constant and will increase only slightly before heat is lost to the surface from all depths.

While no computations of the past changes in temperature below the upper crust have been made because of the numerical work involved, it is possible to obtain a qualitative picture of these changes. Above a depth of 50 km, the temperature-time curves will closely resemble the curves of surface-heat flow against time. In the outer part of the Earth the depth at which the temperature had reached a maximum must have increased with time as shown in Figure 6, but the decrease in radioactive content is so great with increasing depth that even at moderate depths the atomic heat exceeded the heat loss for shorter and shorter periods of time, in spite of the increase in thermal inertia with increase in depth. Thus, at no great distance (possibly 50 km) below the crust, the period of heating must have been confined to the very early life of the Earth as in the upper crust. Therefore in the past, though probably not at present, the upper crust was cooling at the same time as the deeper parts of the crust were heating while at still greater depths cooling prevailed. As time progressed the zone of heating thinned and disappeared as shown in Figure 6. This feature of the Earth's thermal history may have an important bearing on the phenomena associated with the expansion and contraction of the crust.

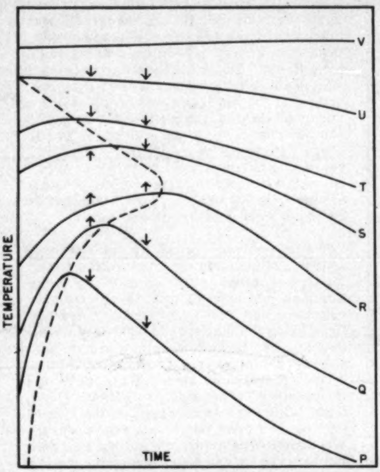

Fig. 6--A qualitative representation of the probable shape of the temperature-time for the Earth at depths of P, Q . . . V curves km; the origin of the time axis corresponds to solidification of the Earth. For purposes of orientation only, the curve for depth V may be considered to hold for all depths greater than the order of 1000 km; curve U might represent the temperature changes 100 to 300 km below the surface, and curve R at the base of the crust

The thermal history of the Earth is one of unexpected complexity; even such a picture as presented in Figure 6 is probably oversimplified because of the many simplifying assumptions made in its derivation.

Parental planet of the asteroids and meteorites--Calculations similar to those discussed in the preceding section have a possible bearing on the disruption of a " parental" planet. According to JEFFREYS [1929] the mass of a planet (average density three gm/cm^3) formed by condensation could not be less than about 0.02 that of the Earth or about twice the estimated mass of the asteroids, a reasonable size in view of the possible capture of much debris by other bodies. Such a parental planet would solidify in less than one-fifth of the time required for the Earth, and the mineral differentiation in the smaller body might be much less complete; that is, its silicate mantle might consist of stony-meteoritic material rather than essentially pure olivine, as it would have to if its composition is indicated by that of the stony meteorites.

From the average composition of the stony meteorites, a meteoritic silicate mantle of a parental planet would be richer in oxides of potassium, sodium, aluminum, and calcium and poorer in magnesia, than a dunitic mantle of the Earth. If the planets were formed by condensation, it might be expected that more of the magnesia, which has a low molecular weight, would escape while still in the gaseous state from the smaller parental planet than from the larger Earth. The lack of water associated with meteorites appears to corroborate this difference, because most of the hydrogen would have escaped from the parental planet. Mineralogically the silicate mantle of the parental planet would be richer in the norm minerals, anorthite, albite, orthoclase, and the pyroxenes. These minerals are associated in the Earth with those rocks into which the radioactivity has been concentrated and which are to be found only in the crust of an Earth which has a dunitic mantle.

Among the possible modes of disruption of a primitive planet, JEFFREYS [1929] mentions explosion as a theory which agrees well with the large eccentricities and inclinations of the asteroids. Jeffreys states "If the planet contained enough radioactive or chemically active matter to heat part of its interior up to the boiling point, the disruptive stresses might come to exceed the small gravitative power of such a mass and permit explosion." Even in a body as small as the proposed parental planet, little or no heat will have escaped from the lower part of its meteoritic silicate mantle. Therefore the temperature rise above the melting point in 2000 million years would be between 1200° and 1800°C and in 3000 million years greater than 1800° to 2700°C. (The decay of the radioelements is exponential. The exact figures for a planet 3000 million years old have not been calculated.) Such temperature rises are more than sufficient to raise the temperature to the boiling point, and the shortest time necessary appears to be about 700 million years which is longer than the minimum time necessary to disperse the resisting medium that would have reduced the large eccentricities of the orbits of the asteroids [JEFFREYS, 1929]. Disruption by this mechanism may not have been confined to a single parental planet, in fact, all celestial bodies of approximately the size proposed above, and with a *meteoritic* silicate mantle, might be expected to have disrupted at some time in the past. The existence of the Earth's moon appears to constitute an exception to this potential instability. The properties of the moon, however, make it unlikely that the moon can be a captured planet or that it could have been formed from any planet other than the Earth, probably by tidal resonance prior to solidification of the

Earth's crust [JEFFREYS, 1929]. In this event its composition must approximate that of the Earth's silicate mantle with an insignificant interior source of atomic heat, although it should possess a crust corresponding to the crust of the Earth with relatively high radioactive content.

Bearing on hypotheses of the origin of the Earth--It appears to be fairly well accpeted that the Earth has a layered structure, and there is much evidence in favor of this view. Proposed Earth models generally agree on the existence of a metallic core surrounded by layers made of minerals composed of oxides and with a decrease in density toward the surface. For this general structure there is a fairly satisfactory explanation on the assumption that the Earth was once fluid. It is also generally agreed by most students of the Earth's present thermal state that the radioelements are concentrated to a very great extent in a thin superficial layer called the crust. As far as is known, no one has offered a reasonably satisfactory explanation of how this layering and differentiation could come about except from a once-fluid Earth. The layered structure, therefore, appears to lend strong support to those theories of the Earth's origin that require the Earth to start off as a liquid [JEFFREYS, 1929], as against theories according to which it formed by the collection of solid particles.

However, as the heat that can escape by conduction from the deep interior of such a sphere within a few billion years is slight, it is evident that if the Earth had been formed by the accumulation of solid matter and if its radioactive material had been distributed uniformly throughout, as would presumably be the case, the temperature in its deep interior would rise because of the atomic heat generated by the radioelements. Given sufficient time and a high enough rate of production of atomic heat, the interior must melt to a fluid Earth that could then differentiate into the layers.

We shall assume that the primordial Earth was solid (crystalline) and that convection was insignificant. Further, we shall assume that the size and mass of the primordial Earth was the same as that of today's Earth and that the total content of the radioelements was that corresponding to the present-day assignment to Adams' earth model given in Table 1, but uniformly distributed throughout the mass. It is found that, if the time that has elapsed since solidification of the Earth is three billion years, the time prior to this necessary to raise the temperature from 0°C and to melt the interior would be of the order of three to four billion years. This would make the age of the Earth and presumably of the solar system about six to seven billion years; which is about the age of the oldest meteorites according to the measurements of ARROL, JACOBI, and PANETH [1942].

In this calculation it is particularly necessary to take into account the exponential decay of the radioelements. For example, the rate of generation of heat seven billion years ago, relative to a given quantity of the radioelements today, was greater by a factor of about 1000 for the uranium-235 series, 40 for the potassium-40, three for the uranium-238 series, and 1.5 for the thorium series.

The time required to accumulate the necessary heat would be less for the models of Washington and Daly or for an entire Earth having the average composition of the meteorites because of the higher total content of radioelements associated with such models. Likewise, had the initial temperature been higher than 0°C, as must surely have been the case for the dust-cloud hypothesis referred to below, the necessary time would have been somewhat less, although it must be borne in mind that the capacity of the assigned amounts of the radioelements to produce sufficient atomic heat to cause melting is largely attributable to the much greater concentrations existing three to seven billion years ago relative to the concentrations remaining today, as explained above.

It should be pointed out that probably the outer portion of the primordial Earth could only be transferred to the melting pot by engulfment. This would be necessary in order to permit the formation of the highly differentiated crust of today.

In spite of some difficulties it would appear worthwhile exploring further the possibilities of this hypothesis because of the increasing evidence making it very difficult to accept the origin postulated by the Jeans-Jeffreys hypothesis and because it appears quite possible that an Earth that originated by the aggregation of small solid parts, as assumed by the new dust-cloud hypothesis, described by WHIPPLE [1948] and PAGE [1948], would, even if it were solid, that is, non-convecting, when first formed, melt and pass through the same stages of separation into layers that has come to be associated with the idea of an orginally liquid Earth.

Acknowledgments--The author is indebted to J. W. Greig of the Geophysical Laboratory, Carnegie Institution of Washington, for suggesting the investigation of the possible melting of an Earth that had originally been created by the accumulation of solid particles.

It is a pleasure to acknowledge the helpful criticisms of Leason H. Adams.

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(Manuscript received December 22, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

ORIGIN AND DEPOSITION OF DEEP SEA SEDIMENTS

Fred B. Phleger, Jr.

(Contribution No. 444 of the Woods Hole Oceanographic Institution)

The upper layers of the Earth's crust are represented in the sea by the sediments which are being deposited from a variety of sources and at a rate which varies from place to place and from time to time. A relatively complete series of sediments should be found in deep-sea deposits if we assume that the present ocean basins have remained approximately as they are now throughout a major part of geologic time. In this discussion the term, deep-sea sediments, is used in a very broad sense but generally to include those deposits which occur on the floors of the large ocean basins below about 2000 fathoms water depth.

The absolute rate of deposition of deep-sea sediment can at present be approached accurately only by radioactive methods such as those used by PIGGOT and URRY [1942]. It should be pointed out, however, that the rates of deposition are variable even in the deep sea and that the same factors affect the rate of deposition in this environment as affect deposition on land: Variations in supply of material, variations in transporting agency, and general geographic position. An additional factor is the marine organic production of the plankton, and this is of measurable importance. Little is known of the origin of the inorganic constituents of marine sediments, but they certainly include what may be called particles of cosmic origin, volcanic ash or dust, non-volcanic terrestrial dust, ice rafted debris, chemical precipitates, and some material which is certainly water transported. The remains of planktonic, and to a very minor extent nektonic and benthonic, organisms are of especial importance in the study of these deposits. It is almost entirely through these organic remains that the past marine environments existing when the sediments were accumulated can be reconstructed, and also the correlation of the deposits from place to place can be accomplished.

It is of interest to review some recent progress in study of deep-sea sediments and some suggestive results which have been obtained. The sequence of marine sediments can be studied by obtaining a cross-section of them in a submarine core, and the optimum condition would be to obtain cores through the sediments to bedrock. It is not possible to obtain such cores at the present state of our experience and with the funds and facilities available. It is generally agreed that a submarine coring tube which operates on the principle of a free fall into the bottom is practical to use and is very successful in terms of results. Such coring tubes are those described by HVORSLEV and STETSON [1946] and by SHEPARD and EMERY [1941]. These obtain a cross-section of the sediments up to a reported five meters in length, with an average length of perhaps 2 1/2 to three meters.

Börje Kullenberg of the Oceanographic Institute of Göteborg has constructed a new type of coring tube which he calls the piston core sampler. This simple gear collects cores up to 20 m in length and makes all other submarine coring tubes in use more or less obsolete in terms of the results obtained. The Kullenberg sampler now is being extensively used to sample deep-sea sediments on the round-the-world cruise of the Swedish Deep Sea Expedition. The writer was a guest of this expedition last summer on their trip through the Carribean and observed the Kullenberg sampler in operation. The core sampler is weighted by several hundred pounds of lead and plunges into the bottom by the momentum gained through the free fall of one meter. This is sufficient to force the tube as much as 20 m into the bottom. There is a piston inside the tube which remains stationary at the surface as the tube falls around it; the apparent effect is to eliminate most of the friction inside the tube and the inside container is completely filled to the depth of penetration. I have seen some of these cores opened immediately after being taken and they represent an excellent cross-section of the material.

This gear opens new possibilities for the study of deep sea sediments and the sequence of events recorded in them. Absolute rates of deposition will have to be determined by radioactive methods. But at least it is possible to establish a long series of events and relative rates of deposition can be fitted into what little is known of the chronology of the Pleistocene. The Foraminifera from one core taken by the Swedish Deep Sea Expedition from the Carribean Sea have been studied by the writer. This core is 1540 cm long and came from a depth of water of about 4800 m. Seventy-five short samples were taken for Foraminifera with the positions of the samples being determined by the layering in the core.

Interpretation of the approximate sequence of surface water-temperature conditions has been based upon the vertical sequence of planktonic Foraminifera faunas in the core. There are at least 12 marked shifts from tropical to very warm temperate water on the one hand to cool temperate to sub-Arctic surface temperatures on the other hand. We do not yet know enough about the ecology of these forms to estimate more specific surface-temperature conditions from the faunas. It is interesting to speculate upon the possible sequence of events represented by the faunas in this core. There is a thick layer of tropical and subtropical faunas between about 800 and 1000 cm. It seems quite possible that this may represent the second interglacial stage of the Pleistocene which we know was as warm or warmer than the present in some places and which is thought to have been at least 200,000 years in duration. If we assume that the rate of sedimentation was approximately uniform throughout the core, then this warm water fauna represents a longer period of time than any other.

If this represents the second interglacial, then the sub-tropical and cool-temperate faunas lying above it must represent the third glacial, the third interglacial, and the fourth glacial. On this interpretation, temperature shifts beneath about 1000 cm may also represent glacials and an interglacial and it is possible that the core may have penetrated most of the Pleistocene.

The pioneer work of Piggot and Urry has shown that the absolute rate of deposition in the deep sea is variable. But if we may generalize on rates of deposition in this environment, we may say that a rate of from one centimeter in 500 years to one centimeter in 1000 years is a fair generalization on the basis of our present knowledge. This core deposited at this rate would take between 750,000 and 1,500,000 years, and this length of time is of the same order as previous estimates for the length of Pleistocene based upon equally deficient data.

The conclusions based upon the circumstantial evidence from this core are very suggestive. The variations in the Foraminifera faunas are very real and represent marked fluctuations in the climate of the sea water in which they lived. Long cores from the sea basins such as this offer the possibility of establishing and correlating a series of marine events of the more recent past from an undisturbed record. If these events can be tied into a carefully placed series of cores on which absolute dates can be fixed by radioactive methods, then we shall know a great deal about the rates of deep-sea sedimentation. Also, we can establish a Pleistocene chronology which should be more complete than any chronology based upon the deposits found on land.

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(Manuscript received June 22, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

FOSSIL POLLEN AND PALEOCLIMATOLOGY

Arthur S. Knox

Abstract

Pollen analysis has been extensively used in Europe not only in studies of postglacial climatic fluctuations, but also in conjunction with investigations of eustatic changes in sea level and isostatic land movements. A beginning in this field of research has been made in several sections of the United States.

In New England late glacial and post-glacial time may be divided into six periods based upon studies of over 50 peat bogs and lake deposits: (1) The pre-Boreal, a cold wet period during the disappearance of the ice sheet in eastern North America, with spruce - scrub pine forests predominating. (2) The Boreal, a dry period with pine forests marking the transition from the spruce to deciduous and hemlock forests. It is the logical marker for the beginning of postglacial time. (3) The Atlantic period, wetter and warmer than the present during which regional differences in the forest flora began to appear. (4) The sub-Boreal, a period of relative dryness. During this time the post-glacial climatic optimum occurred when heat loving plants reached their maximum development. (5) The sub-Atlantic, which witnessed a deterioration to wetter and cooler conditions as shown by a general increase in spruce. (6) Recent, a somewhat warmer and drier period.

The exact dating of these periods of climatic change offers somewhat of a problem in this country. They are, however, similar in position and character to dated climatic fluctuations found in European peat bogs, and evidence from several sources suggests that they were at least approximately contemporaneous in northeastern North America and northwestern Europe.

The value of these studies was demonstrated when an Indian fishweir was uncovered in Boston, Massachusetts a few years ago, 15 ft below present sea level. Pollen analyses of peat and sediments in which the weir was entombed indicated that it was built soon after the climatic optimum about 4000 years ago. Tentative rates of sedimentation and submergence of this section of the Atlantic Coast were determined by these studies.

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(Abstract received May 7, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust.)

DIURNAL VARIATIONS IN THE ACCELERATION OF GRAVITY

E. A. Eckhardt

The daily rhythm of the tides is familiar to everyone who has lived or vacationed near tidewater. The forces which produce the tides are due to the attraction of the Moon and the Sun. These same forces act over the land masses where their effects are less obvious and less well known. Formulas and tables are available from which the tide producing forces may be computed for any place and any moment of time. The forces so computed for any point on the Earth's surface over any period of time would correspond to the variations with time of the acceleration of gravity at that point, if the Earth were unyielding. The actually observable variations will depart from those computed because the Earth yields in response to the tide-producing forces.

The tidal or diurnal variation of gravity has a maximum range of approximately 0.3 mgals over the Earth's surface, or roughly three parts in ten million of total gravity. To obtain data for a curve of the tidal variation of gravity, it is necessary to measure gravity differences to one part in one hundred million or better. In addition, the measurement must be accomplished in a very short time, measured in minutes. The first instrument approximately fulfilling these requirements was developed by Tomaschek and Shaffernicht, especially for tidal measurements. It was a recording and non-portable or observatory type of instrument, depending for its sensitivity on close approach to an instability configuration. The calibration therefore required constant surveillance and detailed consideration at each point of the recorded curve. As discussed by LAMBERT [1933], the amplitudes of the tidal-gravity observations by Tomaschek and Shaffernicht are substantially less than those computed for a rigid Earth. For a non-rigid Earth, they should be greater. Measurements made by WYCKOFF [1936] indicated a much closer agreement between computed and observed results, the amplitudes on the whole being slightly greater but at intervals also less than the predicted values. In the light of our own measurements, the results of Tomaschek and Shaffernicht definitely seemed anomalous. In the spring of 1939, Gulf Research & Development Company had 12 gravimeter parties scattered over a latitude range of about 47° and a longitude range of over 127°. It was arranged that these parties make simultaneous observations at their respective locations over an extended week end. Observations were made at half-hour intervals. The instruments used were all of the same model known to us as the standard Gulf Torsion Spring Gravimeter. It is essentially a weight supported by a spring. The spring consists of a spirally wound ribbon so designed that any extension results in a corresponding rotation. The angular rotation is amplified by a multiple reflection optical system. The material of the spring has a minimum temperature coefficient of elasticity, and the instrument is thermally insulated and the temperature is thermostatically maintained constant to within several hundredths of a degree. The least count of the instrument is something under 0.01 mgal.

A preliminary report of the simultaneous observations made by us was presented at the Washington meeting of the International Union of Geodesy and Geophysics in September 1939. The results were not otherwise published, although it has always been our intention to give this subject further study. For various reasons this has not been possible until now.

Our furthest south station was at Santa Barbara in eastern Venezuela, where the computed tidal variation had a computed average double amplitude of 0.22 mgals and that observed was 0.25 mgals. Our furthest north station was at Varde, Denmark. Here the computed average double amplitude was only 0.076 mgals, or about one-third of that at Santa Barbara. The observed average double amplitude was 0.094 mgals.

All our data show the progressive decrease of amplitude with latitude going north, including the extremes already mentioned. Instead of discussing the data for the individual stations, it will suffice to point out that out of 128 turning points recorded, 100 gave observations exceeding those computed, with 18 cases nearly equal and only ten cases in which the observed values were less than those computed. It seems to us established therefore that the substantial deficit of observed amplitude reported by Tomaschek and Shaffernicht, if real, must have only local sifnificance and may not be conceded to have more general validity. It seems to us that our measurements establish beyond any reasonable doubt that the tidal variations of gravity which may be measured are, in general, somewhat larger than those computed from the formulas for the tide producing force. This is what one would expect.

The Tomaschek and Shaffernicht measurements also show a very pronounced phase lag between the computed and observed maxima, ranging from 14° to 97° for the various components into which their curves were analyzed. Our curves suggest strongly that any phase lag must be small. To determine it, it would have required making our observations more frequently than the half-hour intervals we used.

At one of our stations at Brownfield, Texas, we happened to have two instruments available and both were read throughout the two and one-half days devoted to the tidal observations. The two instruments had somewhat different drift curves. The comparison of the data for the two instruments with each other and with the computed variations strengthens our feeling that the observed departures are real and that some of its details may have local significance. The two independent observations made at Brownfield, Texas, while exhibiting some minor differences, show a convincing similarity in the curves exhibiting the differences between the observed and computed values, to which I refer as the gravity departures.

As sheer speculation, one is tempted to imagine the data to suggest that the Earth yields to the tidal forces as would an ice field, which is composed of blocks of various sizes. The whole field responds to the tides in a general way, but with appreciable departures from the general mean for the various individual blocks which are more rigid with respect to the tidal forces than the mass as a whole. These blocks may be tilted or lifted more or less than the Earth's surface over wider areas, and there might be consequent mass displacements which influence the observable diurnal variation of gravity.

It is at least scientifically intriguing to visualize a program of extensive and systematic measurements of the diurnal variations of gravity. Instruments designed for prospecting, such as we used, are not particularly suitable for such work. Since our observations had to be made visually continuously over a period of days, three observers were required for each instrument, or a total of 36 for the localities at which we made observations. These observers were not equally skillful, however enthusiastic. A systematic program would require recording instruments. Such a program would involve measurements at many places over appreciable periods of time. That would require many instruments, and they should therefore be relatively inexpensive. They should cost no more than about a tenth of what prospecting gravimeters cost. They should perhaps be readable to something less than 10^{-8} gamma. It might be found, however, that this extra sensitivity is not practically usable. It is not usable in general prospecting.

We have given considerable thought to the design of such an instrument. However, the intervention of the War and the subsequent insatiable world appetite for oil and more oil have prevented us, up to now, from following any further avocational inclinations.

Finally, it is a pleasure to acknowledge the enthusiastic cooperation of members of my staff, too numerous to mention individually, in making the instrument observations and reducing them to usable form.

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(Manuscript received December 17, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

ATTENUATIONS AND AMPLITUDES OF SEISMIC WAVES

Norman Ricker

In researches previously published [RICKER, 1940, 1941, 1943, 1944, 1945] , the author has shown theoretically that the disturbance created by the detonation of a small charge of explosive in a homogeneous and isotropic Earth assumes a certain characteristic form, shaped by the Earth's absorption spectrum for elastic waves. This form and its variants have been called seismic wavelets. According to the theory, when observed at a great distance from the explosion, the wavelet should assume a symmetrical form. Figure 1 shows the form of the displacement, and represents the motion of the trace of a mechanical seismograph as it would record the motion, the mechanical seismograph being properly designed so as to avoid distortion. Figure 2 shows the form of the velocity-type wavelet and represents the motion of the trace of an electrical seismograph which is velocity responsive and which also has been designed to avoid distortion. Figure 3 shows the form of the acceleration-type wavelet. In most of the experimental work, with a velocity responsive electrical seismograph, blasting caps and charges of but a few pounds of dynamite have been used. In clay and shale the velocity form of Figure 2 has been observed, as shown on specimen seismograms given in one of the early papers [RICKER, 1940]

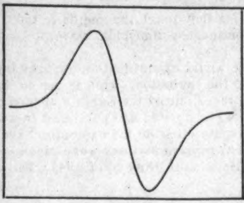

Fig. 1--Displacement type wavelet $U(u|100)$

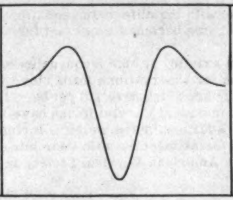

Fig. 2--Velocity type wavelet $V(u|100)$

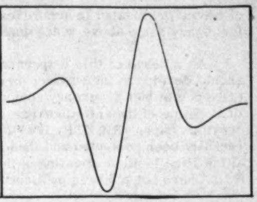

Fig. 3--Acceleration type wavelet $W(u|100)$

As long as we do not approach too close to the site of the explosion, the following laws of propagation of the wavelet are predicted by the theory.

(1) It is the center of the wavelet which travels with the classical velocity of sound in the medium, this velocity being given by the square root of the elasticity over the density. This law holds for all three forms of Figures 1, 2, and 3.

(2) The breadth of the wavelet, measured in seconds, say, is proportional to the square root of the propagation time of its center, the law holding for all three forms of Figures 1, 2, and 3.

(3) The amplitude of the displacement-type wavelet of Figure 1 decays as the wavelet advances through the Earth in such a manner that the amplitude is inversely proportional to the 4/2 power of the travel time of its center. This is for spherical waves spreading out from the center of disturbance.

(4) The amplitude of the velocity-type wavelet of Figure 2 decays in such a manner that its amplitude is inversely proportional to the 5/2 power of the travel time of its center. This also is for spherical waves spreading out from the center of disturbance.

(5) The amplitude of the acceleration-type wavelet decays in such a manner that its amplitude is inversely proportional to the 6/2 power of the travel time of its center. This again is for spherical waves spreading out from the center of disturbance.

It has been found that in certain clays and shales the absorption which shapes these wavelets and which should account for their laws of propagation is due to internal friction of the type caused by viscosity. The viscosity which causes this absorption, however, is not the viscosity of continuous, unidirectional deformation under plastic flow, but is of an entirely different and smaller order of magnitude, being of the order of 10^7 gm/cm sec or even lower.

As a result of the studies made of these seismic wavelets it has become quite evident that the classical equation of wave propagation

$$\nabla^2\chi = (1/c^2)\partial^2\chi/\partial t^2$$

is sufficient properly to explain the nature of actual seismograms and that, for a better description of the propagation of seismic disturbances it is necessary to make use of a modified equation of wave propagation

$$\nabla^2[\chi_1 + (4\eta/3\rho c^2)(\partial\chi_1/\partial t)] = (1/c^2)(\partial^2\chi_1/\partial t^2)$$

which, in its form for plane waves, was first given by Stokes in 1845. The above equation is for dilatational disturbances. For shear disturbances the equation is

$$\nabla^2[\chi_2 + (\eta/\rho\bar{c}^2)(\partial\chi_2/\partial t)] = (1/c^2)(\partial^2\chi_2/\partial t^2)$$

These equations of wave propagation may be written

$$\nabla^2[\chi_1 + (1/\omega_0)(\partial\chi_1/\partial t)] = (1/c^2)(\partial^2\chi_1/\partial t^2)$$

and

$$\nabla^2[\chi_2 + (1/\bar{\omega}_0)(\partial\chi_2/\partial t)] = (1/\bar{c}^2)(\partial^2\chi_2/\partial t^2)$$

If we write $W_0 = 2\pi f_0$ and $\bar{W}_0 = 2\pi\bar{f}_0$, then f_0 and $\bar{f}_0$ are called the transition frequencies for the dilatational and shear disturbances, respectively. These frequencies are important in the study of waves propagated in accordance with the differential equation, for they mark the region in the frequency scale above which dispersion becomes important for continuous, sinusoidal waves.

As a result of this dispersion existing at high frequencies one would expect that an asymmetry should develop in the wavelet form for observations made close to the explosion. This is due to the presence of high frequency components which have not yet been filtered out by the Earth's absorption. Some of the anticipated asymmetrical wavelet forms have been computed and published in a previous paper [RICKER, 1943]. Additional asymmetrical forms quite close to the explosion have recently been computed and these forms together with their integral representations were given at the Rice Institute meeting of the American Physical Society in Houston in November 1947. These forms have not yet been published.

In measuring pressures developed in the Earth due to explosions it is customary to measure the height of the pressure peak first arriving. Since the pressure follows the velocity form we may compute the anticipated pressure as a function of the distance from the explosion by determining the magnitude of the first peak of the velocity form. If we do this we find that for reasonably large distances the pressure should be inversely proportional to the 5/2 power of the distance from the explosion, but as the site of the explosion is approached the curve becomes abnormally steep due to the asymmetry of the wavelet forms so that the pressure at these close distances varies inversely as the third, fourth, or even higher powers of the distance.

As the site of the explosion is approached it appears that the form described above as the displacement form, approaches an impulse type motion at the explosion itself. That is to say, at the site of the explosion the Earth moves away from the origin and back to the origin in a very short interval of time. The forms observed experimentally and described above were created by small charges planted at a shallow depth. One would expect that if the charge were so planted that a pressure could be built up and sustained so that the initiating pulse became a unit step function instead of an impulse, the resultant wavelet forms should be the integrals of the forms described above.

Very close to the explosion there is most likely to be plastic flow of the Earth, and since this entails a much larger coefficient of viscosity, it is to be expected that the resulting abnormally greater absorption near the explosion should cause departures from the laws of wavelet propagation given above for short distances from the explosion.

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(Manucript received June 25, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

THE USE OF ROCKBURSTS IN THE STUDY OF CRYSTAL STRUCTURE

John H. Hodgson

Since 1939 a series of rockbursts have been occurring in the gold mines at Kirkland Lake. These rockbursts recommend themselves as a source of energy for the investigation of crustal structure, since the mechanism of energy release has been shown to be similar to that of earthquakes. A seismograph has been installed at Kirkland Lake to time the occurrence of the bursts and a series of seismograph stations has been installed along a traverse towards Ottawa. By changing the location of these stations as bursts are recorded, a refraction profile will be built up. While the program is in operation the stations serve also as additional stations of the Canadian chain.

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(Abstract received June 25, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust.)

EFFECT OF PRESSURE ON PHYSICAL PHENOMENA IN THE CRUST

Roy W. Goranson

The physical properties of materials as tabulated in handbooks and as used in our everyday life refer to measurements made in the environment which surrounds us, namely at atmospheric pressure (slightly more than one bar at sea level).

If we were able to move down into the interior of the Earth with our testing equipment we would find the environmental pressure increasing at a rate of about 30,000 bars for each 100 km of descent. We would find, moreover, that under these conditions our testing apparatus would record very different physical behavior from that which might be anticipated from observations made at the surface under our normal atmospheric environment.

Forty odd years ago F. D. Adams of McGill University made some rather simple experiments to demonstrate a few of these differences. He surrounded his test specimens with plaster or steel jackets to increase the confining pressure. His technique has been reproduced on some marble test samples which are shown in Figure 1. The rectangular prism was compressed with very little confinement, the middle specimen with heavier confinement, and the third with a steel jacket. No evidence of shear planes are present in the third specimen. This technique is not a very satisfactory one because the confinement cannot be controlled or measured with any certainty.

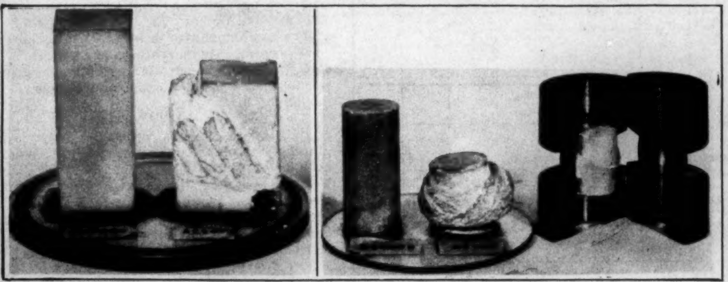

Fig. 1--Tests of specimens of marble

The following illustrative examples are excerpts from an experimental program of eight years ago which was interrupted by the War before it could be completed and published.

The apparatus used is described elsewhere [GORANSON, 1940]. The hydrostatic confining pressure, which is transmitted to the specimen by a thin hydrocarbon fluid, and the added unidirectional thrust can be varied independently of each other.

In Figure 2 is shown a series of steel test specimens. This material has a high tungsten content (known as high speed tool steel) and was heat treated until glass hard. From left to right are an original untested sample, a pile of small fragments and powder which are the remnants of a

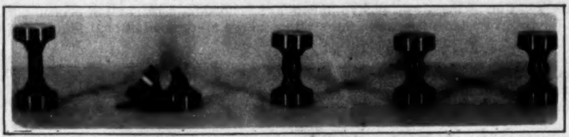

Fig. 2--Tests of high tungsten steel

specimen compressed under normal atmospheric conditions until it ruptured. No flow was observable. The next three were compressed at successively higher confining pressures.

A plot of differential stress, that is, uniaxial compressive load intensity minus the hydrostatic pressure, versus hydrostatic confining pressure, is given in Figure 3. The lower curve represents the yield point and the upper curve the ultimate or rupture strength of the steel. At one atmosphere of confinement no shortening by yield was apparent, at 5000 atmospheres the specimen shortened 36 per cent before failing by shear, between 10,000 and 15,000 atmospheres the specimens shortened 47 per cent before failing by shear. The resistance to shear failure increases more rapidly with increasing confining pressure than does the yield point. As a matter of fact, the increase in yield point and in melting point with increase of confining pressure is very similar.

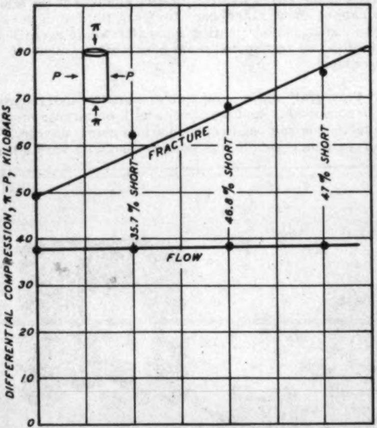

Fig. 3--Relationship of flow and fracture to differential in pressure

The material of Figure 4 is hard-drawn tungsten rod. The initial unstressed state is denoted by the center specimen. The one on the left was compressed at one atmosphere confining pressure and shattered with no observable preliminary yield. The one on the right was compressed in an 18,000 atmosphere environment. The barreling, as a result of flow, is clearly discernible (the straight upper portion was supported inside of a cylinder during the test).

Cylinders of quartz were also tested under similar conditions but no yield was discernible even under confining pressures of 25,000 bars. The crushing strength, however, had increased seven fold, as shown in Figure 5b. Pyrex glass, which is also very brittle under ordinary compressional tests, showed a similar type of increase in crushing strength, as shown in Figure 5a. Under confining pressures of 17,000 or more bars, however, definite yield by flow was observed.

Cylinders of carboloy, which had been used as pistons in the inner second stage of my cascaded pressure apparatus, would sometimes explode into fine dust on removal after a run indicating that flow and recementation had occurred so as to relieve the internal stress distribution resulting from the external loading. When these external forces were removed the material was no longer at ease with the new environment and could only relieve the frozen-in stresses explosively.

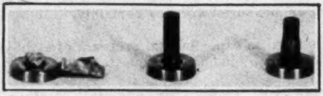

Fig. 4--Tests of a hard-drawn tungsten rod

It therefore appears that under sufficiently high confining pressure, abetted by increasing temperature, materials will reach their yield point and relieve non-uniform stress distributions by flow long before slowly applied stresses can accumulate sufficiently to reach the rupture strength of the material.

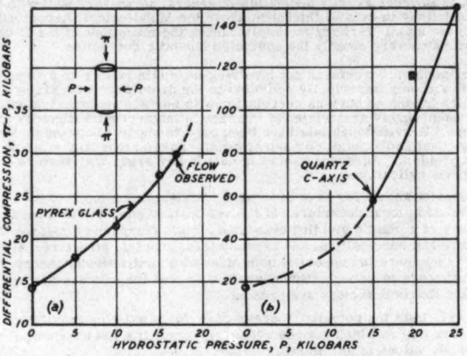

Fig. 5--Tests of cylinders of quartz

It seems improbable, therefore, that deep focus earthquakes, which occur at about 380 km depth, and thus under confining pressures in excess of 100,000 bars and at high temperatures, could result as a consequence of sudden failure. A more probable hypothesis would be to assume these earthquakes result from abrupt volume changes caused by temperature-pressure phase transitions.

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(Manuscript received December 10, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

GRAVITY ANOMALIES AND THE NATURE OF THE EARTH'S CRUST

George Prior Woollard

Abstract--Gravity anomalies appear to constitute a reliable method of studying crustal structure and composition in any one area. However, because of uncertainties in base-station values of gravity, the inaccuracies of early gravity observations, and the magnitude of the Browne Correction on early marine surveys, the extent to which the present anomalies can be used for inter-continental and inter-oceanic comparisons as to the nature of the crust is questionable. Taken as a whole, the continental areas appear to be in isostatic equilibrium. Notable exceptions, however, occur in connection with some of the major sedimentary basins and with mountains and valleys developed as horsts or grabens. Part of the anomalous gravitational field in these areas can frequently be accounted for by the local geology, but the major portion of the anomaly usually appears to be related to deep-seated crustal effects.

In the oceanic areas, existing isostatic gravity anomaly values indicate extensive regions which are characterized by anomaly values of one sign. This suggests either a marked variation in the shape of the Earth from that assumed, or variations in crustal composition and structure. The occurrence of extensive areas having isostatic anomalies of one sign (positive) in the smaller marine areas as the Gulf of Mexico and the Mediterranean Sea favor the latter explanation. The occurence of lenticular belts of positive and negative anomaly values in association with island arcs is believed to reflect a local thinning and thickening of the crust.

Variations in Bouguer gravity anomalies in general agree very well with seismic evidence for variations in crustal thickness above the Mohorovicić Discontinuity. In connection with the major Tertiary mountain ranges the magnitude of the Bouguer anomalies approximates very closely the computed isostatic correction.

Most major geologic structures not involving mountain ranges or piedmont sedimentary basins are only imperfectly reflected in the gravity values. Major regional gravity anomalies having no obvious correlation with surface geologic features are quite common and probably are related to changes in intra-crustal lithology and crustal structure. Gravitational data have been used to support both the thesis of a plastic yielding crust and a rigid, competent crust. Taken alone they offer no unique solution to this problem. Accessory seismic data in key areas may resolve these apparently ambiguous indications.

Introduction--Gravity anomalies have been used to determine a more exact figure for the shape of the Earth, to study local departures of the vertical, to support the theory of isostasy, to support both the theory of a plastic yielding crust and a rigid competent crust for the Earth, to study crustal structure and composition, and to study local geologic structure. Since gravity data are frequently used as evidence in support of both sides of a controversial theory it is evident that either their significance is imperfectly understood or that the data are subject to question. At this time it appears that both factors are present.

HEISKANEN [1941] lists ten potential sources of error in existing gravity anomaly values and to this list there are others which may be added. A composite list of potential sources of error in gravity anomaly values is as follows.

(1) The accuracy of the base station used
(2) The accuracy of the equipment used
(3) The accuracy with which the observations are tied to base stations
(4) The accuracy of determination of the elevation of the station
(5) The particular gravity formula used
(6) The method of isostatic compensation assumed
(7) The density of the crust assumed
(8) The accuracy of the available topographic and bathymetric charts used in reducing the data
(9) The accuracy of the method used in applying terrain corrections
(10) The accuracy of the reduction tables used
(11) The accuracy of the evaluation of the effect of pitch and roll on marine observations
(12) The accuracy with which instrumental drift and diurnal changes are evaluated
(13) The effect of undulations of the geoid

The probable error involved for each item of the above list varies from only one or two to ten milligals or more. Heiskanen reports that the mean error of 47 older stations in the Carpathian region of Europe which were reoccupied was 18 mgal with some stations differing as much as 35 mgal or more. Similarly, he points out that relative to Potsdam many national base stations appear to be seriously in error. For example, values in India are believed to be about 15 mgals high whereas those in Switzerland are believed to be about 12 mgal low.

The average error that can be attributed to the instruments used varies with the individual instrument, type of instrument, and method of timing used. Observations with standard geodetic pendulums operated by competent personnel and using radio-time signals to rate clockwork chronometers appear to have an over-all accuracy of about ± two mgal. On the other hand, a special quartz pendulum apparatus has been built, which when used with a crystal chronometer, has an accuracy of about 0.1 mgal. Precision-type gravimeters may be read with an accuracy of about 0.01 mgal. As most published gravity data were obtained using standard bronze or invar geodetic pendulums timed with clockwork chronometers, most of the data available are probably subject to ± two mgal error.

The effect of errors in elevation is approximately 0.6 mgal for ten feet of elevation, but since elevation control on the whole has been good, the error from this cause is generally small. Since most data are now reduced using the International Spheroid of reference the effect of changes in gravity formula is largely obviated. It should be noted though, that some geodetic surveys do not use the International Spheroid in their reductions. The combined difference for observations in India due to error in base value and difference in gravity formula used may be as much as 30 mgal in some areas.

The effect of using different systems of isostatic compensation will vary with the terrain of an area. In flat low-lying country the difference in results for using different systems of compensation is small; in high rugged country there may be a difference of 20 mgals or more in the anomalies. The effect of the density value used in the reductions will vary with the local geology. Normally a mean density of 2.67 is used for the crust of the Earth in reducing gravity data for geodetic purposes. The fact that the actual density to considerable depths may differ markedly from this value may give rise to errors as large as 20 mgals or more in some areas.

The effect of inaccuracies in evaluating the gravitational effect of the surrounding terrain will vary with the local area and may be only a few tenths of a mgal or four or five mgals or more in very rugged areas. The errors incorporated in the method of considering terrain or the type of reduction tables used are usually very small, less than one mgal. The result of omitting the effect of pitch and roll on submarine gravity measurements will vary with the state of the sea and with the depth at which the observations are made. Errors resulting from this cause, however, may amount to 20 mgal or more. The undulations of the geoid may have an effect as great as eight mgal but over most of the Earth this effect will only amount to about ± three mgal.

The total potential error under certain conditions may therefore be quite large, even larger than many of the anomalies obtained that are believed to have geodetic and geologic significance. Fortunately the probable error from all causes for any group of observations relative to a common base (omitting the effect of pitch and roll on marine observations, and local density changes on land observations; and assuming good equipment and competent operation personnel) is seldom more than five or six mgal. With these limitations in mind we may proceed to examine anomalies on a world-wide basis and attempt to evaluate them in terms of the nature of the Earth's crust.

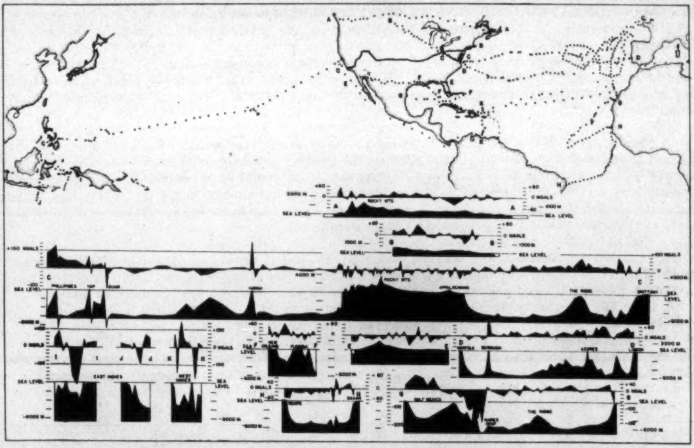

Fig. 1--Isostatic gravity anomaly and topographic profiles of North America, the Pacific Ocean, and the Atlantic Ocean

The continents versus the oceans--In Figure 1 are shown several gravity traverses crossing the Pacific Ocean, North America, and the Atlantic Ocean. These traverses are for the most part based on pendulum data. They have not been corrected for undulations of the geoid or changes in density, and there is a possibility of about six mgal shift in the base value between the various areas. All of the data were isostatically reduced according to the Pratt theory of isostasy.

Section AA across southern Canada shows that although the Rocky Mountains are essentially in isostatic equilibrium the mid-continent region except for the area immediately north of the Great Lakes appears to have been overcompensated in computing the reductions as evidenced by the predominant positive anomalies. The area of negative anomalies north of the Great Lakes may reflect a residual crustal downwarp from the Pleistocene ice load since it coincides in position with the area which was presumably the center of the Pleistocene ice cap. If this interpretation of this gravity low is correct, then on the basis of the probable density contrast of 0.5 between crustal and subcrustal material and the anomaly present the region will have to rise about another 4000 ft to be in isostatic equilibrium.

Section BB runs from the Minnesota Shield area out to Cape Cod. This section shows that there are large areas that are very much out of isostatic equilibrium. There is a prominent, positive anomaly area in the exposed Canadian Shield and a marked negative anomaly area in the Appalachian region. The high values may reflect a local thinning of the granitic crust since it can not be attributed to the densities of the surface rocks, and the low values may reflect a residual "root" to the Appalachian Mountains.

Section CC which crosses both the Atlantic and Pacific Oceans as well as North America indicates (1) whereas the western Pacific between Guam and the Philippines appears to be undercompensated, the balance of the Pacific area is essentially in isostatic equilibrium and does not differ markedly from the continental area of North America; (2) in contrast to the Pacific area the Atlantic area is characterised by marked positive anomalies and appears to be undercompensated. (All the oceanic sections in Figure 1 are based upon the revised data of F. A. Vening Meinesz and differ in some areas from the sections that he originally published.)

If the above relations are interpreted in terms of lithology they are at variance with the seismic data which indicate that the rocks underlying the Pacific Ocean transmit at a higher velocity and are probably denser and more basic than those underlying the Atlantic Ocean. An alternative explanation is that the Earth's shape is actually that of a triaxial ellipsoid and as proposed by HEISKANEN [1938] has a major axis running between 25°W and 155°E. The absence of positive anomalies at the Pacific axial position, however, raises some doubt as to this explanation. The difference in the width of the anomalies in the oceanic and the continental areas is believed to reflect the greater number of stations over the continental area. This is borne out by the observations of VENING MEINESZ [1934] which showed that when two stations were located close together large changes in anomaly over short distances were the rule rather than the exception.

In connection with the bathymetric troughs usually associated with island arcs it is seen from Section CC and also Sections II, JJ, and KK that the gravity anomaly profiles across all of these are very similar although in the western Pacific the mean anomaly value becomes progressively more positive as the Philippine Islands are approached. This change in the mean anomaly value from Guam to the Philippines appears to be too abrupt and displaced too far to the west to fit the effect from the triaxial Earth proposed by Heiskanen.

Alternative explanations are that the observations are in error, or that it is significant that the break in the anomalies correlates with the position of the Andesite Line which is a structural boundary between Asia and Australia and the Pacific Basin. The volcanic islands to the west of the Andesite Line are of an andesitic nature while those to the east of the line are basaltic. In addition, the islands to the west of the line have the geologic structure of folded continental ranges whereas those to the east are volcanic peaks with no evidence of folding or orogeny. Since the Andesite Line also follows the belt of island arcs, bathymetric troughs, and earthquake epicenters, there may be tectonic significance in the gravity relations observed. However, the data of a single traverse are too few to do more than suggest a relationship which should be explored further.

Section DD crosses the North Atlantic slightly to the south of Section CC and the two are in substantial agreement. One significant feature on this traverse is the large positive anomaly (+ 95 mgal) associated with Bermuda. This is similar to that observed at Hawaii on Section CC. By way of contrast the anomaly associated with the Azores appears to be essentially zero. The explanation of this difference is not obvious since on Section CC at the crossing of the Mid-Atlantic Ridge on which the Azores are located there is a pronounced positive anomaly.

Local positive anomalies such as are shown at Hawaii and Bermuda are also found on the island of Cyprus. (The average isostatic anomaly on Cyprus is + 105 mgal with a maximum of + 173 mgal [MACE, 1939]. In the Hawaiian Islands the average anomaly is + 114 mgal with maximum of + 209 mgal [DUERKSEN, 1943].) These occurrences may be explained on the assumption that in these areas, isostatic compensation is achieved on a regional basis and that the rigidity of the crust is such that it can sustain large localized loads over long periods of time.

Section EE is across the southern United States and shows that the area is in isostatic equilibrium except in the Mississippi Delta area where positive anomalies occur in juxtaposition to a marked gravity low which is apparently associated with the Gulf Coast geosyncline, a downwarped area filled with sedimentary material.

Section FF which runs from New Orleans across Florida and the Bahamas is distinguished primarily by the occurrence of a marked positive anomaly area in the Gulf of Mexico. A similar relation is obtained in the Mediterranean Sea area. In neither area is there an obvious explanation for the relations observed.

Section GG running from Brownsville, Texas, to Africa shows that the positive anomalies in the Gulf of Mexico shown on Section FF are quite general. The wide negative anomaly area shown results from the traverse following along the axis of the bathymetric trough in front of the West Indian Island arc. Section KK is a cross section of this same feature. In the Atlantic Ocean area of Section GG isostatic equilibrium is more nearly approached than on the northern traverses, and the Mid-Atlantic Ridge appears to be essentially in isostatic equilibrium.

Section HH is a traverse across the south Atlantic from Recife, Brazil, to Dakar, Africa. On this section it is seen that the amplitude of the anomalies is not only much less than those obtained on the other Atlantic traverses but the anomalies tend to be negative rather than positive. The only departure from an otherwise undistinguished gravity field is found in the sharp "high" and "low" associated with a submarine peak and trough on the Mid-Atlantic Ridge.

Sections II, JJ, and KK are comparative sections across the marked gravity lows associated with the bathymetric troughs in the East Indies and the West Indies. (The West Indian section is somewhat different from that originally published by HESS [1938] in that it incorporates later data taken by Vening Meinesz.) In both areas the significant features are the same; the negative anomalies are on the convex side of the arcs and are associated with the ocean deeps, and the positive anomalies occur on the concave side of the arc usually in association with the islands...

From a study of the above sections in Figure 1 which incorporates most of the available gravity data for the oceanic areas of the world, the following relations appear to exist: (1) Most of the major topographic irregularities of the Earth's surface are isostatically compensated, (2) many of the small topographic features are not isostatically compensated, (3) there are large anomalies which can not be correlated with topography, and (4) sweeping generalizations such as "the oceans are characterized by positive isostatic anomalies and the continents by negative anomalies" are not justified by the available data.

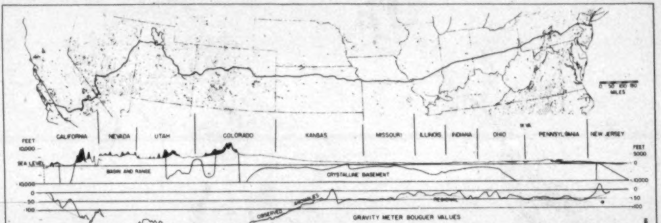

Fig. 2--A gravimetric and magnetic traverse of North America

Gravity anomalies within a continental area--In Figure 2 are shown the results of a transcontinental gravity traverse across North America along which observations were made at about seven-mile intervals. The Bouguer anomalies which reflect the actual mass distribution vary from about -50 mgal over the eastern half of the country to over -300 mgal beneath the Rocky Mountains, and the magnitude of the Bouguer anomalies appears to be directly related to the elevation of the stations. If a smooth curve is run through the base of the Bouguer anomalies and compared with a similar curve based on the amount of the isostatic correction computed for some 50 key stations along the traverse, it is found that the two are in very close agreement. The crust as a whole along this traverse of the continent therefore seems to be essentially in isostatic equilibrium. The extent to which local isostatic compensation is obtained cannot be ascertained from this section because of the masking effect of the large anomalies present which are related to unknown mass distributions within the crust.

On the basis of the available seismological data, the large negative Bouguer anomalies probably represent, for the most part, changes in crustal thickness. However, whether they represent a change in the thickness of the upper layer, or the intermediate layer, or of both cannot be stated. In Southern California BYERLY [1939] and GUTENBERG [1943] report that the upper granitic layer in this area apparently maintains the same thickness beneath the Sierra Nevada as is found over the rest of Southern California (18 km) but there is an increase in thickness of about 30 km in the intermediate layer beneath the Sierra Nevada.

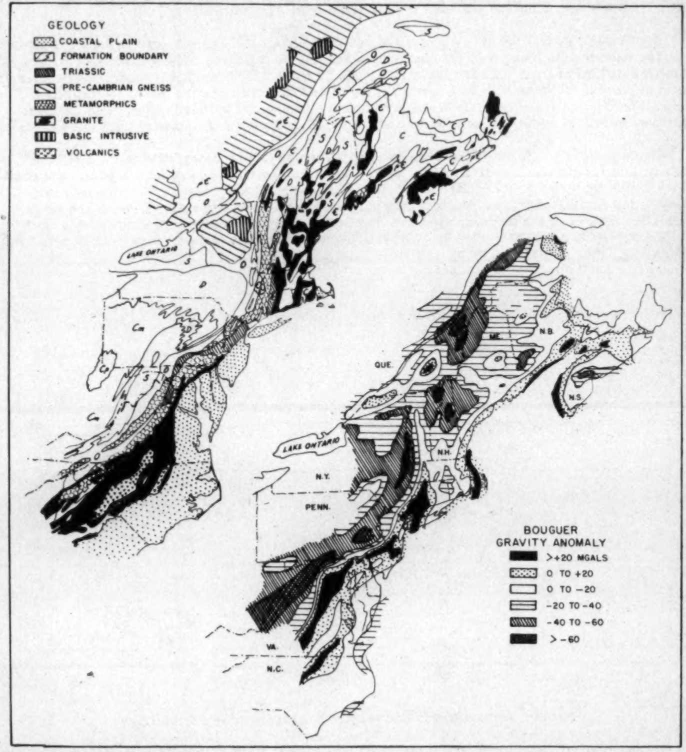
GEOLOGY
COASTAL PLAIN
FORMATION BOUNDARY
TRIASSIC
PRE-CAMBRIAN GNEISS
METAMORPHICS
GRANITE
BASIC INTRUSIVE
VOLCANICS
LAKE ONTARIO
LAKE ONTARIO
QUE.
ME.
N.B.
N.S.
N.H.
N.Y.
PENN.
VA.
N.C.
BOUGUER
GRAVITY ANOMALY
>+20 MGALS
0 TO +20
0 TO −20
−20 TO −40
−40 TO −60
>−60

Fig. 3--Bouguer gravity anomaly and geologic map of Eastern Seaboard of the United States

In addition to the general completeness of isostatic equilibrium exhibited on this traverse and the evidence for variations in crustal thickness, the section shows conclusively that there are large isostatic anomalies (± 30 mgal) which for the most part cannot be correlated with known geological features. This situation exists despite the fact that actual density changes were ignored and a standard density value of 2.67 was used in all the reductions. The principal exceptions where the effect of local density changes materially affected the anomalies were in the Green River Basin, over the Uncompaghre Uplift, and in the Great Salt Desert.

These large isostatic anomalies which cannot be correlated with known geologic or topographic features are believed to reflect for the most part intra-crustal changes in lithology resulting either from intrusive or tectonic disturbances.

Areal gravity studies in Eastern North America--In Figure 3 is shown an isoanomaly Bouguer map of the eastern seaboard from the Maritime Provinces of Canada through Virginia. This map is based on about 5000 observations and is believed to approximate closely the actual anomalous gravitational field present in the area. The parallel arrangement of positive and negative anomaly belts and the close similarity in trend between the gravity anomalies and the geology are immediately apparent. That the anomalies do not reflect the densities of the surface and shallow geologic formations, however, is indicated in the occurrence of the belt of most pronounced positive anomalies in the Coastal Plain region of low density sediments. The negative gravity anomaly belt follows the line of overthrusts on the eastern side of the folded Appalachian Mountains.

South of Long Island the positive gravity anomaly belt appears to be split by a series of minor gravity lows along its axis which appear to correlate with the position of Paleozoic granites. A correlation between gravity lows and granites is also obtained in northern New England and in the Maritime Provinces of Canada. MILLER [1946] has postulated that the granites of New Brunswick and Nova Scotia must extend to a depth of at least ten km to explain the gravity anomalies associated with them. In 1948 D. C. Skeels in a personal communication also called attention to a similar relation in Finland where younger granites intruded into older granites are characterized by local negative gravity anomalies.

The gravity low in the White Mountains of New Hampshire can be explained in part at least on the basis of isostatic compensation. The Green Mountains on the other hand occur along the axis of the positive anomaly belt, and appear to have no root. In the Adirondack Mountain region of New York the heavy basic igneous rocks present exert a greater effect upon the anomalies than the compensation for the topography as is evidenced by the local increase in the anomaly values.

The most significant gravitational feature along the Eastern Seaboard, however, is that the anomalies occur as parallel positive and negative belts paralleling the geologic tectonic axes. This pattern is suggestive of flexures in the crust as proposed by VENING MEINESZ [1940] to account for the similar gravity pattern obtained in connection with island arcs. However, since there is only one major positive and negative anomaly pair upon which are superimposed lesser gravity structures the data may be interpreted as reflecting an overriding and underthrusting of two sheared ends of a faulted crust as postulated by GUNN [1944, 1947]. The parallel lesser anomaly bands result under either hypothesis from lithologic changes due to intrusions along shear planes. Under the assumption of the sheared crust as proposed by Gunn the overrriding block would be on the seaward side. This is in agreement with the geologic facts in that the sediments of the Appalachian geosyncline were derived from the east and the overthrusts were from the east to the west. However, it should also be pointed out that the similarity in gravity and seismic patterns and geologic and petrologic associations between orogenic belts on land and those associated with island arcs is so pronounced that a genetic relationship between the two appears quite probable, as proposed by HESS [1939]. The general relations associated with island arcs, which are in large measure duplicated on land in connection with major mountain ranges, have been outlined by GUTENBERG and RICHTER [1945] and are as follows.

(1) An oceanic trench (deep)

(2) Belt of shallow-focus earthquakes and negative gravity anomalies which occur on the concave, landward, side of the deep, sometimes in association with a ridge on which there are folded non-volcanic islands

(3) A belt of positive gravity anomalies with large earthquakes of shallow to intermediate focal depth

(4) The principal structural arc with active or recent volcanoes; earthquakes with focal depths of 100 km and decreasing gravity anomalies

(5) A second arc of older volcanoes with earthquakes having focal depths of 200-300 km.

(6) A belt where there are deep-focus earthquakes (300-700 km deep)

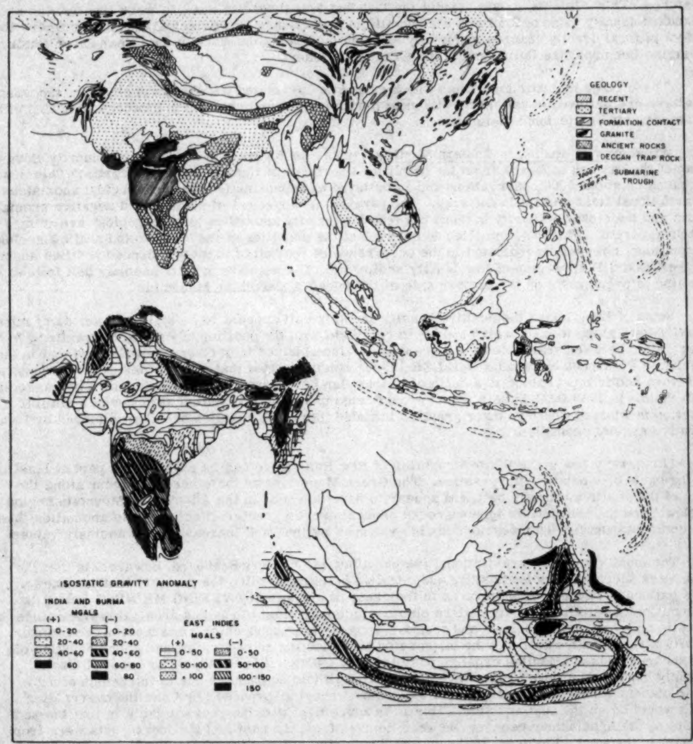
GEOLOGY
RECENT
TERTIARY
FORMATION CONTACT
GRANITE
ANCIENT ROCKS
DECCAN TRAP ROCK
3000 fm
SUBMARINE TROUGH
ISOSTATIC GRAVITY ANOMALY
INDIA AND BURMA
MGALS
(+)
0-20
20-40
40-60
60
(−)
0-20
20-40
40-60
60-80
80
EAST INDIES
MGALS
(+)
0-50
50-100
. 100
(−)
0-50
50-100
100-150
150

Fig. 4--Isostatic gravity and geologic maps of southern Asia

To this list should be added the presence of peridotite intrusions in the most highly deformed zone which HESS [1937] believes represents basic magma that entered at the time of the primary downbuckle of the crust as postulated by VENING MEINESZ [1940] or at the base of the shear plate as postulated by GUNN [1947]. If continental mountain ranges and the island arcs are genetically related, then the orientation of the pattern relative to the land masses is probably significant from the standpoint of the study of permanence of the continents and the ocean basins. Present island arcs with only one or two exceptions face away from the land toward the ocean basins. The continental mountain ranges face away from the sea in that the overthrusts are generally toward the land areas.

Gravity relations in southern Asia--In Figure 4 isostatic anomalies for southern Asia are shown. Marked negative anomaly areas occur in the Indus-Ganges sedimentary basin at the foot of the Himalaya Mountains, in southern India, and in Burma. On the basis of seismic relations which exist between Java and Burma (the progression from shallow-focus to deep-focus earthquakes) the Burma negative area is a continuation of the negative anomaly belt associated with the East Indian island arc., Certainly the negative anomalies of Burma cannot be explained by the density of the local geological formations. In connection with the negative anomalies of the Ganges

Valley, EVANS and CROMPTON [1946] showed that about 75 mgal of the negative anomaly shown could be removed by assuming this sedimentary basin extended to a depth of 11 km and was filled with sediments having a specific gravity of about 2.5. This correction, however, which assumed the sediments were nearly all coarse elastics with essentially no compaction, only removed about one-half of the original anomaly. Additional evidence that the observable geology is not the dominant control of the anomalies here is shown by the negative anomaly values found over the high-density Deccan traps in western India. Similarly in the case of the Cudapah Basin in eastern peninsula India a negative anomaly is found although the basin is filled with Cambrian sediments with a density of about 2.7. On the basis of the above relations, SKEELS [1940] concluded that the negative anomalies associated with the sedimentary basins of India were the result of local downwarps of the crust as a whole, rather than the density of the material filling the basins.

These same relations appear to obtain in Turkestan (not shown in Figure 4) as isostatic anomalies of -60 mgals are obtained in the Ferghana Basin.

In addition to this evidence of crustal warping in southern Asia, there is the broad positive anomaly belt crossing peninsula India which has been postulated by GLENNIE [1932] to be the result of a broad crustal upwarp of about nine km.

In southern Asia there is therefore gravitational evidence in support of crustal warping over and above that associated with mountain ranges. In addition it appears from seismic data and the available gravity data in the East Indies and southern Burma that island arc-type structures might carry continuously from ocean areas on to continental areas.

Bouguer anomalies in Europe--In Figure 5 is shown the areal distribution of Bouguer anomalies as observed over a part of Europe. It is seen that the same relations noted in North America are also found here. Large negative Bouguer anomalies are associated with the Apennine Mountains, the Dinaric Alps, the Alps, the Bohemian Massif, the Carpathian Mountains, and the Caucasian Mountains, and extensive areas of positive anomalies occur in the north German Baltic Coastal Plain, the intermontane Roumanian Basin, the Aegean Sea, the Mediterranean Sea, and the Black Sea areas. Although some of the anomalies undoubtedly reflect changes in lithology, the majority of the anomalies shown appear to reflect variations in over-all crustal thickness which are substantiated in large measure by seismological data. In contrast to the situation previously cited in Southern California where the increase in crustal thickness is accommodated mostly in the intermediate layers, in Europe the increase in crustal thickness appears to be accommodated largely in the upper granitic layer. A summary of data on variations in crustal thickness in Europe as prepared by GUTENBERG [1943] is shown in Table 1.

Table 1--Variations in crustal thickness (after Gutenberg)

Area	Thickness of layers		Depth to D (Mohorovicic discontinuity)
	Granitic	Intermediate	
	km	km	km
NW Europe	30	10	40
Swiss Alps	25-30	20-25	50
N. Alps	35	20-25	55-60
Carnic Alps	40	20-25	60-65
Yugoslavia	15	25	40

An interpretation by SALONEN [1932] of the Bouguer anomalies observed in the Swiss Alps in terms of changes in crustal thickness indicates a thickening of about 16 km beneath the Alps, which value agrees fairly well with the seismic determinations given in Table 1.

Although the major mountain ranges of Europe such as the Alps appear to be isostatically compensated, there are large areas of negative isostatic anomaly associated with the piedmont sedimentary basins. Conditions are therefore similar to those observed in Asia in that the mountains appear to be in equilibrium and the basins are out of equilibrium by amounts which are too large to explain on the basis of the local density distribution without involving crustal warping.

Areas not in isostatic equilibrium--An examination of data from various parts of the world indicate that certain geologic tectonic conditions appear to be characteristically out of isostatic equilibrium. In Figure 6 comparative gravity profiles are shown on the same scale for a few of these.

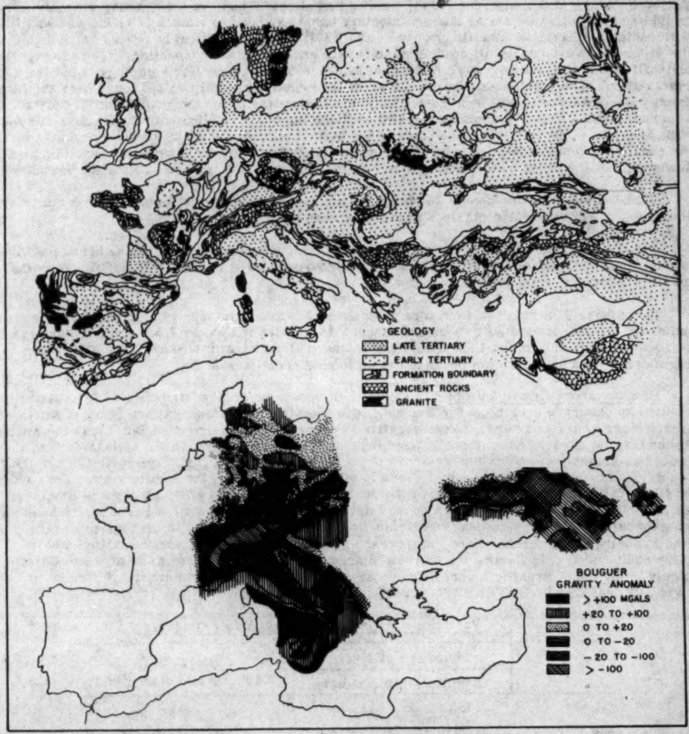

Fig. 5--Bouguer gravity anomaly and geologic map of Europe

The first section is across the Po Basin at the foot of the Carnic Alps in northern Italy. On this section both Bouguer and isostatic anomalies are shown. The negative gravity anomaly associated with the Po Basin and the Apennine Mountains is not removed after isostatic compensation although that for the Alps is removed. SKEELS [1940] has shown that in order to explain this large negative isostatic anomaly associated with the Po Basin on the basis of the density of the sediments present and with reasonable allowances for compaction at depth, the basin would have to extend to a depth of at least 29 km. If the seismic evidence for the thickness of the crust in Yugoslavia can be taken as normal for southern Europe (15 km for the granitic layer) this would mean an actual displacement at depth of the intermediate and deeper crustal layers in the Po region of at least 15 km.

The second section shows the isostatic anomaly obtained over the Nares Deep in the West Indies and is typical of the gravity anomaly associated with island arcs. As postulated by VENING MEINESZ [1934] this anomaly pattern is related to a downbuckle of the upper granitic crust which displaces denser substratum at depth. The third section presents the isostatic anomaly obtained in crossing a very old mountain system which has been postulated to have had the same type of origin as an island arc. This section, taken in Virginia, shows that the axis of the negative belt, seen in plan in Figure 3, underlies the Blue Ridge Mountains and the line of overthrusts on the east

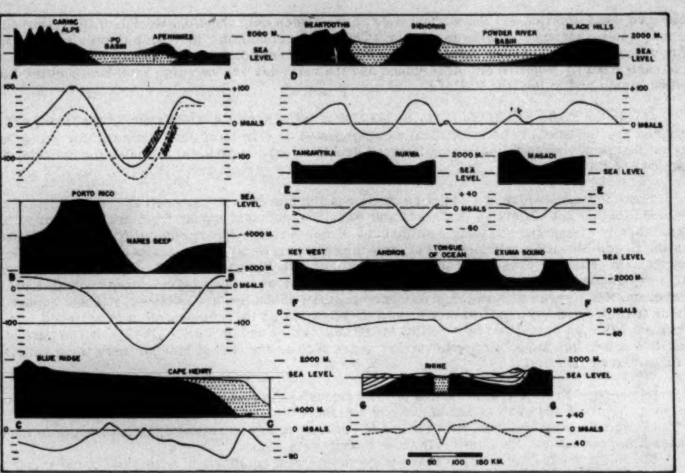
CARNIC ALPS
PO BASIN
APENNINES
2000 M.
SEA LEVEL
A
A
100
+100
0
0 MGALS
-100
-100
PORTO RICO
SEA LEVEL
NARES DEEP
-4000 M.
-8000 M.
B
B
0
0 MGALS
-100
-100
BLUE RIDGE
CAPE HENRY
2000 M.
SEA LEVEL
-4000 M.
C
C
0
0 MGALS
-80
BEARTOOTHS
BIGHORNS
POWDER RIVER BASIN
BLACK HILLS
2000 M.
SEA LEVEL
D
D
+100
0
0 MGALS
TANGANYIKA
RUKWA
2000 M.
SEA LEVEL
MAGADI
SEA LEVEL
E
E
+40
0
0 MGALS
-60
0
KEY WEST
ANDROS
TONGUE OF OCEAN
EXUMA SOUND
SEA LEVEL
-2000 M.
F
F
0
0 MGALS
-60
2000 M.
SEA LEVEL
RHINE
2000 M.
SEA LEVEL
G
G
+40
0
0 MGALS
-40
0 50 100 150 KM.

Fig. 6--Geologic structures not in isostatic equilibrium; AA, Po Basin, North Italy; BB, West Indian Island Arc; CC, Appalachian Mountain System, United States; DD, Big Horn Mountains and Basins, Montana and Wyoming, United States; EE, Rift Valleys, Africa; FF, Bahama Islands; GG, Rhine Graben, Germany

side of the Appalachian Mountains. The crest of the positive anomaly belt which occurs in the present piedmont area has been broken by a series of minor lows which correlate in position with carboniferous granite intrusives. In addition to the gravity pattern which parallels the geologic structural trends, it appears significant that other features such as the location of a belt of active shallow-focus earthquakes along the axis of the negative gravity belt persist, and there are volcanics and peridotite intrusions in association with the positive anomaly belt as is the case with island arcs. The significance of the marked gravity high shown on this section just inside the edge of the Continental Shelf is not definitely known, but it would appear to be related to intracrustal lithology since there is a large magnetic anomaly in the same area.

The first section on the right-hand side of the figure is for the Big Horn Mountain region of Wyoming where the structurally high mountain regions are characterized by positive isostatic anomalies, and the structurally low basin areas are characterized by negative isostatic anomalies. Both CHAMBERLIN [1935] and the writer [WOOLLARD, 1938] tried to account for these anomalies on the basis of the known geologic section using actual densities determined from well cores and outcrops. The maximum effect due to the geology appeared to be about 15 mgal and the general anomaly picture remained unchanged. In this area both the mountains and the sedimentary basin are out of isostatic equilibrium, and the gravity anomalies appear to be related to crustal displacement. If the curves of SKEELS [1940] are used to calculate this displacement a differential movement of the mountain and basin blocks of the order of ten km is indicated.

The second and third sections are for two of the Rift Valleys in Africa where BULLARD [1936] reports there is essentially no accumulation of sediments in the valleys. Here similar relations are obtained as in the Big Horn region in that there are negative isostatic anomalies in the structurally low valleys and positive anomalies over the structurally high areas. The crustal displacement according to Bullard appears to be of the order of eight to ten km.

The fourth section is for the Bahama Banks where essentially land-locked ocean deeps are characterized by pronounced negative isostatic anomalies relative to the land areas. HESS [1937] has postulated that the anomalies here might be explained on the basis of the high-density dolomites lying above the general level of the ocean floor, but even if a density differential of 0.2 is assumed between the dolomites and the normal rock of the region only about one-fourth of the observed anomaly can be accounted for on this basis. The final section is for the Rhine Graben in Germany for which Bouguer anomaly values are shown. On this section the structurally low valley is characterized by negative anomaly values and the relations are the same as obtained in the Big Horn region and in the Rift Valleys.

From the above examples it can be concluded that either the dimensions of the features examined are too small to be isostatically compensated or else local compressional forces in the crust last long enough to maintain displaced crustal blocks in artifical attitudes for considerable periods of time.

Some features which are out of isostatic equilibrium such as the central areas of depression beneath former ice sheets as in Canada and Fenno-Scandinavia appear from geologic data to be isostatically rising and attaining equilibrium. Other features apparently out of equilibrium such as the Green Mountain anticline in Vermont may be superficial features superimposed by coincidence on the axis of larger crustal structures at depth, or they may be the actual surface expression of a primary displacement of the crust at depth. In the case of isolated island peaks such as Hawaii, Bermuda, and Cyprus with which pronounced positive anomalies are obtained, it would appear that these features are too small to be compensated locally and that compensation is achieved on a regional basis as postulated by VENING MEINESZ [1934] and by GUNN [1947]. In the case of the positive anomalies associated with marine areas such as the Gulf of Mexico there is no satisfactory explanation at this time.

Summary--From a consideration of the various areas examined the following relations concerning the crust and gravity anomalies can be made:

(1) No sweeping statements can be made concerning any universal difference between the crust beneath the oceans and the continents from gravity data alone.

(2) On the whole, the crust appears to be in isostatic equilibrium, with most of the major topographic irregularities compensated. Where lack of compensation is indicated, as over broad areas of the North Atlantic, changes in the depth of compensation used will bring the results more into line. Physically this might be interpreted as indicating a change in the depth at which plasticity is reached within the crust, or a change in the thickness of the granitic crust.

(3) Locally there are marked departures from isostatic equilibrium associated with mountain ranges, sedimentary basins, island arcs, rift valleys, and graben. It is impossible to tell from the gravity data whether these tectonic forms are the results of crustal downbuckling induced by straight compressional forces or by convection currents or whether they involve over- and underthrusting of rigid competent crustal material. That the crust is not a uniform concentric layer is indicated by seismic observations, and variations in crustal thickness of 30 km appear to be quite common. This thickening and thinning of the crust appears to be of the same general order of magnitude as that indicated by the major variations (>100 mgal) in the Bouguer anomalies.

(4) The explanation of smaller anomalies (± 30 mgal) within the continental shields and ocean basins that are not associated with structural units such as mountains, island arcs, sedimentary basins, and other features, is not definitely known, but presumably these anomalies are related for the most part to intracrustal changes in lithology.

(5) That the crust is rigid enough to sustain superimposed loads such as volcanic mountains and delta deposits without appreciable isostatic yield is indicated by the positive anomalies associated with islands such as Bermuda Cyrpus, and Hawaii, and with the Mississippi delta. That the crust has yielded to the broader loading of ice caps is indicated by the residual negative isostatic anomalies and the succession of raised beaches in Fenno-Scandinavia and northern Canada, and that it still is rising in these areas is indicated by present tidal data.

On the basis of the gravity data and subsidiary seismic data, the crust therefore appears (1) to yield to large superimposed loads such as ice caps but not to local loads such as volcanic islands; (2) to vary in thickness so as to compensate for the change in load occasioned by major topographic irregularities; (3) to yield to accumulative loads such as are found in sedimentary basins; (4) to respond to compressional forces by failure in shear and possibly flexure along lines of apparent weakness near the continental boundaries; (5) to vary in apparent thickness over broad areas in a random manner; and (6) to have a heterogeneous lithographic composition.

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(Manuscript received July 20, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

SEA LEVEL CHANGES ALONG THE COASTS OF THE UNITED STATES IN RECENT YEARS

H. A. Marmer

In the determination of mean sea level as a geophysical datum it has been recognized that sea level derived from one year of observations will differ from that derived in another year, primarily because of the disturbing effects of wind and weather. For this reason it has been considered desirable to base a mean sea-level determination on a number of years of observations.

In recent years it has become evident that in spite of the relatively large variations in sea level from year to year due to wind and weather, systematic tide observations can bring out the slow secular changes taking place in the relation of land to sea. The United States Coast and Geodetic Survey now has systematic tide observations at a number of places along the coasts of the United States for periods of 20 years and more. And these provide quantitative data regarding changes in the relative elevations of land to sea in recent years.

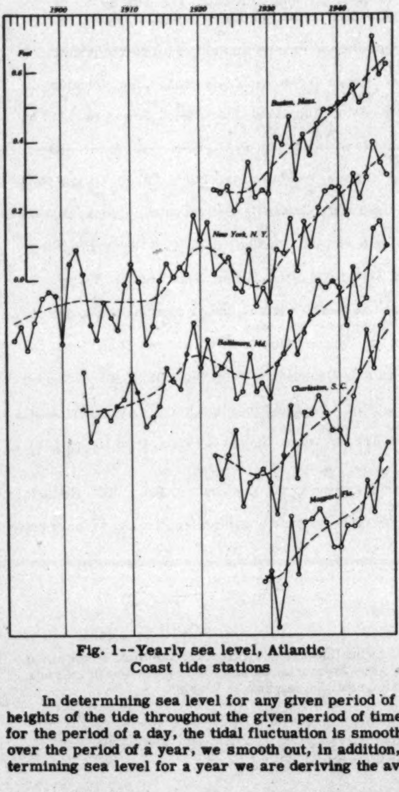

Fig. 1--Yearly sea level, Atlantic Coast tide stations

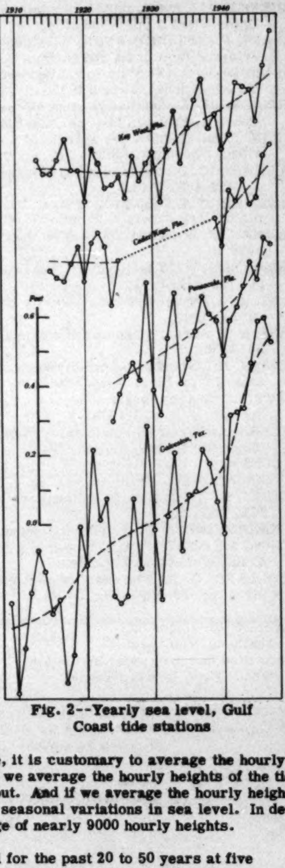

Fig. 2--Yearly sea level, Gulf Coast tide stations

In determining sea level for any given period of time, it is customary to average the hourly heights of the tide throughout the given period of time. If we average the hourly heights of the tide for the period of a day, the tidal fluctuation is smoothed out. And if we average the hourly heights over the period of a year, we smooth out, in addition, the seasonal variations in sea level. In determining sea level for a year we are deriving the average of nearly 9000 hourly heights.

In Figure 1 are shown the yearly heights of sea level for the past 20 to 50 years at five stations along the Atlantic Coast, from Boston, Massachusetts to Mayport, Florida. From one year to the next, sea level is seen to vary from a few hundredths of a foot to as much as two-tenths of a foot. These variations are brought about primarily by the variations in wind and weather from one year to another. A simple method of smoothing out these meteorological variations is by the method of moving means. The dashed-line curve associated with the diagram for each station was drawn by deriving moving means and then passing a more or less smooth curve through these means.

Taking New York, for which we have the longest series of observations, it is seen that from 1893 to about 1930 there was relatively little change in sea level. The apparent rapid rise from 1914 to 1920 and subsequent rapid fall to 1929 can be shown to be chance effects of wind and weather and not real changes in sea level. But from 1930 to the present time, sea level has been

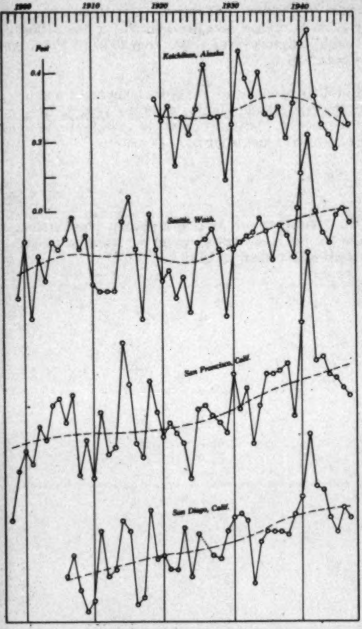

Fig. 3--Yearly sea level, Pacific Coast tide stations

rising steadily so that in 1947 it was 0.36 ft higher than in 1930, or at the rate of approximately 0.02 ft per year. From 1893 to 1930 the rise was only 0.13 ft or at a rate only one-sixth that since 1930.

The data for Baltimore present similar features. There was little change to 1930, and a steady rise from that year to the present time. Since 1930 the rise is 0.35 ft, practically the same as at New York, while for the 27-year period from 1903 to 1930 it was 0.12 ft or at a rate about one-fifth that since 1930.

The other three stations furnish definite information only for the period subsequent to 1930. Since that year the rise for Boston is 0.32 ft; for Charleston 0.40 ft and for Mayport 0.32 ft. The average for the five stations is 0.35 ft, so that the rate of rise for this stretch of the Atlantic Coast is almost exactly 0.02 ft per year since 1930.

For the thousand-mile stretch of coast from Boston to Mayport, the conclusion is inescapable that there has been a relative subsidence of the Coast during the past 17 years at the rate of about 0.02 ft per year. It is important to stress that this is a relative subsidence, for from these sea level data alone we cannot determine whether it is an absolute subsidence of the Coast, an absolute rise in sea level or a combination of the two.

For the coast on the Gulf of Mexico, data are available for the four stations shown in Figure 2. Key West shows much the same features as the Atlantic Coast stations. There was little change in sea level to 1930, after which the rise is at the rate of a little less than 0.02 ft per year. At Cedar Keys there was a break in the observations from 1925 to 1938. But there are strong indications of little change in sea level prior to 1930, and since 1939 the rise has been at the rate of about 0.02 ft per year.

At Pensacola, sea level appears to have risen more or less steadily since the beginning of the observations in 1924. In the intervening 23 years sea level has risen 0.40 ft, that is at a rate a little less than 0.02 ft per year. However, in the latter half of this period the rate of rise is greater than in the first half.

At Galveston, there has been a steady rise of sea level since the beginning of the observations in 1909. The total rise during the 38 years of observations is 0.82 ft or at an average rate of a little over 0.02 ft per year. From 1909 to 1930 the rate of rise was about 0.015 ft per year, while from 1930 the rate has been 0.03 ft per year or double the rate prior to 1930.

So far as the Gulf Coast as a whole is concerned, the indications are for greater relative subsidence since 1930.

For the Pacific Coast of continental United States, Figure 3 gives the yearly sea levels for San Diego, San Francisco, and Seattle. For these three stations the rise of sea level shows an acceleration in the more recent years, though the year 1930 is not so clear-cut a dividing point as is the case on the Atlantic Coast. Using that year for convenience, sea level shows a rise of 0.10 ft for San Diego, 0.13 ft for San Francisco and 0.11 ft for Seattle. For this stretch of the coast therefore, the average rise in sea level since 1930 is at the rate of about two-thirds of a hundredth of a foot per year, or at a rate only one-third that found on the Atlantic Coast.

The diagram for Ketchikan, Alaska is of interest in showing that there are regions along the coast of the continent which are not exhibiting the general rise of sea level noted for the Atlantic, Gulf, and Pacific Coasts of the United States. At Ketchikan, sea level rose from 1919 to 1935, then remained stationary until 1940 and since then has been falling.

The varying rates of rise or fall of sea level at different places along the shores of a continent obviously point to differential movements of the continental block. With the results of systematic tide observations available throughout the world, it will be possible to determine what part of such movements is due to a general rise of sea level and what part is crustal.

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(Manuscript received June 4, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 23, 1948 as part of the Alfred C. Lane Memorial Symposium on the Earth's Crust; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

SUPERCOOLING AND CRYSTALLIZATION

J. Ross Heverly

Abstract--The supercooling of droplets of water under conditions of cooling rate and pressure that bracket those found in nature has been quantitatively studied. Also the temperature at which water vapor in ordinary atmospheric air will crystallize to form snow crystals has been determined. The spontaneous freezing point is presented in a graph as a function of droplet size. Such varied water sources as tap water, distilled water, mountain stream water, and water condensed from the atmosphere were used. A photomicrograph of formvar replicas of the snow crystals formed is presented, and the size distribution and relative number of the accompanying supercooled water droplets is given in tabular form, as a function of temperature. The results of the supercooling and crystallization investigations indicate that the current precipitation theories need modification. A new concept of the initiation of precipitation is presented.

Introduction--The literature pertaining to the supercooling and crystallization of water contains many results of significance, but also contains many contradictory claims and much that is conjecture and unwarranted theorizing. As early as 1858, MOUSSON [1858], reported that water in the form of a fog could be supercooled, but as late as 1929 a text [HUMPHREYS, 1929] with the physics of the air ignored the existence of supercooled droplets. OLTRAMARE [1879] reported that water at a temperature as low as -19°C could be vigorously agitated without its freezing, but many modern physics texts still claim quiescence necessary.

Since 1900 there have been two noteworthy contributions to the study of the freezing point of water en masse. MIERS and ISAAC [1906] used the term "spontaneous freezing point" for the temperature at which water freezes in the absence of introduced ice particles. However, they reported -1°.6 C to -2°.0 C as the spontaneous freezing point. DORSEY [1938] found spontaneous freezing points, temperatures at which water froze "in the twinkling of an eye," of -10°C to -20°C. Dorsey found the freezing point to depend upon the size of the largest "mote" present in the water. His work disproved the necessity of quiescence for extreme supercooling. Dorsey's work also refuted the ice-crystal-molecule theory because the results were independent of the rate of cooling to the spontaneous freezing point. Both of these investigations used sizable volumes of water (a number of cubic centimeters). Previous to the present study [HEVERLY, 1948] there had been no systematic investigation of the supercooling of droplets of water.

According to the BERGERON [1935] -FINDEISEN [1938] theory, the formation of snow crystals on ice crystal nuclei in the clouds precedes any large amounts of precipitation. This theory was defended by STICKLEY [1940] but his article indicates that by means of the indirect synoptic approach it is about as easy to disprove the Bergeron-Findeisen theory as it is to confirm it.

Some experimental evidence was set forth by STÜVE [1931] regarding the type of condensate that would be formed with various types of nuclei at temperatures below 0°C, and, in the last few years, there have been quite a number of experiments dealing with crystallization of atmospheric water vapor. Experimental evidence as to what happens in very rapid expansive cooling of ordinary air was set forth by CWILONG [1947]. In Wilson cloud-chamber studies, Cwilong found crystallization to occur in ordinary air at -32°C ± 1°C. WEICKMANN [1945] studied the formation of atmospheric ice crystals, and collected samples of them from the various types of clouds. SCHAEFER [1946] found crystallization to occur in an open-top cold chamber at about -35°C. More recently SCHAEFER [1948] has refined this estimate to -39°C ± 0°.1 C except for a relatively few crystals which he observed at as high a temperature as -35°C.

Method and apparatus--In the present crystallization investigations [HEVERLY, 1948], cooling and expansion, in simulation of the natural atmospheric processes, were employed. The highest temperature at which snow crystals formed was determined by preserving, in replica form [SCHAEFER, 1941], the condensates produced at successively lower temperatures.

A general view of the apparatus used in the current study is given in Figure 1. The temperature was measured by means of thermo-couples and a combination potentiometer-galvanometer

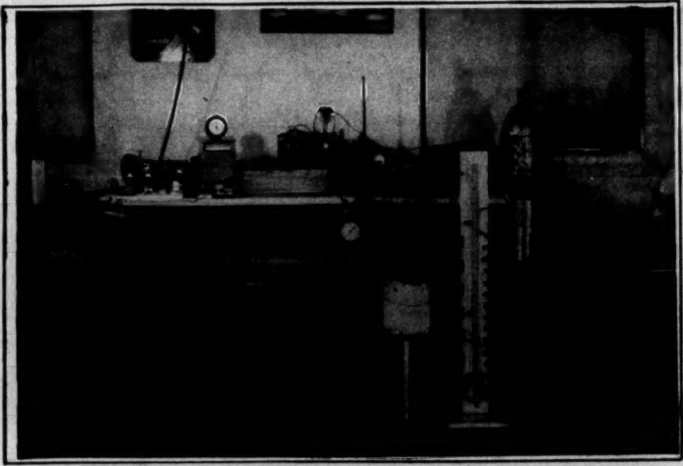

Fig. 1--General view of crystallization apparatus

method so that the readings could be taken very rapidly as the test cell was cooled. The test cell was submerged one-half inch beneath the surface of the rapidly circulating bath of a cryostat. The cryostat (see Fig. 1, right center foreground) was originally designed by SCOTT and BRICKWEDDE [1931] but the design was modified by the writer in order to provide working space in the bath liquid without the use of gears.

In the supercooling investigations, droplets of water were placed in the test cell and their spontaneous freezing points determined for cooling rates varying from one to 20°C /min and for pressures varying from 986 mb to 8.7 mb. It is apparent that this pressure range extends from approximately standard atmospheric pressure to less than the lowest pressures that would be encountered by natural water droplets. Vertical velocities of about 20 m/sec are the highest that have ever been measured in cumulonimbus clouds [STANLEY, 1943] . On the basis of saturated-adiabatic cooling this would correspond to about seven°C /min. Even assuming velocities, calculated from the size of hailstones, of twice this speed, the cooling rates employed in this investigation bracket those to which a cloud droplet may be subjected in nature.

The support of the droplets consisted either of the tiny thermocouple junction, used in the measurement of temperature, or of a small piece of wax paper. The droplets were present either singly or in groups.

The cell was arranged for microscopic observation and care was taken that heat from the microscope lamp illumination, or radiation or conduction from the microscope, would not affect the drops. A Leitz Ultrapak microscope was used which allowed illumination from above. The light beam was sent through a heat-absorbent glass and the lucite immersion cap of the microscope objective.

In the crystallization investigations, the condensates, produced in the test cell, were forced into a formvar solution by a small differential pressure. Both a jet method [KINOSHITA and UCHIYAMA, 1932] and an aspiration method [HAGEMANN, 1936] were used in obtaining this differential pressure. On evaporation of the solution, perfect formvar replicas of the condensates remained.

Results--The water droplets investigated ranged in diameter from 0.05 mm to 1.1 mm (50 to 1100 microns). The source of the water was widely varied; it included (a) water obtained from the air by sublimation on the clean outer surface of a beaker filled with dry ice; (b) water condensed from steam that had been produced by the boiling of tap water; (c) boiled tap water; (d) boiled tap water which had been cooled to nearly 0°C; (e) distilled water; (f) one per cent sucrose solution; and (g) one per cent sodium chloride solution. In cases (c) through (g) the droplets were formed by forcing the water through an atomizer. The results in all of these cases, (a) through (g), were found to lie along the same curve, namely, a curve showing the relationship between spontaneous freezing point and droplet size (see Fig. 2). Therefore, the spontaneous freezing point of the droplets depended on the droplet size but, within the wide range of sources enumerated, the spontaneous freezing point of the droplets was independent of the source. (An example of a source-dependent spontaneous freezing point is given toward the end of this paper.) It is also evident from Figure 2 that the data obtained were independent of the type of support.

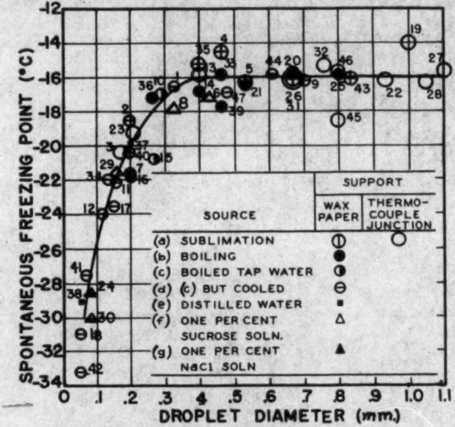

Fig. 2--Spontaneous freezing point of water droplets as a function of droplet size

In addition to the simultaneous use of visual observation, the freezing of droplets having diameters larger than 0.3 mm (300 microns) could be determined from galvanometer deflection. The heat of fusion from the freezing droplet was sufficient to raise the temperature of the tiny thermocouple junction, even though the droplet on the wax paper was about two millimeters away from the thermocouple junction.

The spontaneous freezing point of water droplets was found to be independent of cooling rate and independent of pressure. For example, cooling rates for five droplets which froze at between -27°C and -31° C varied from seven to 18°C/min. As an example of the lack of a pressure effect, it may be observed that droplets numbered 43, 25, 9, 31, 44, 21, 35, and 34 (numbers listed are in order of decreasing size of droplet) in Figure 2 fall along the same freezing curve as the other droplets, although these droplets were frozen at pressures as low as 8.7 to 60 mb whereas most of the other droplets were frozen at atmospheric pressures corresponding to much lower altitudes.

For drop diameters from 1.1 to 0.4 mm (1100 to 400 microns) the spontaneous freezing point was found to be practically constant at -16.0° C, the standard deviation being 0.5° C. Even mountain stream water was found to freeze spontaneously at -15.4° C (0.5 mm drop, not included in Figure 2). Below 0.4 mm, 400 microns, the spontaneous freezing point decreased to a value of -30°C for a diameter of 0.06 mm, 60 microns. Below 0.4 mm (400 micron) the rate of decrease of freezing point with decreasing size was found to vary inversely with the drop size.

The drop sizes investigated vary from the size of small rain drops to the size of the larger cloud droplets. The smallest drop diameter investigated was 50 microns. This is within the range of the results reported by DIEM [1942] who found nimbostratus cloud-droplet diameters to range up to 70 microns. Köhler, in 1923, found a mean diameter of 40 microns (radius 20 microns) for fog and nimbostratus. Diem reports that information regarding drop-size distribution is not available for cumulus or cumulonimbus clouds but it is presumed that they are considerably larger than in stratus-type clouds.

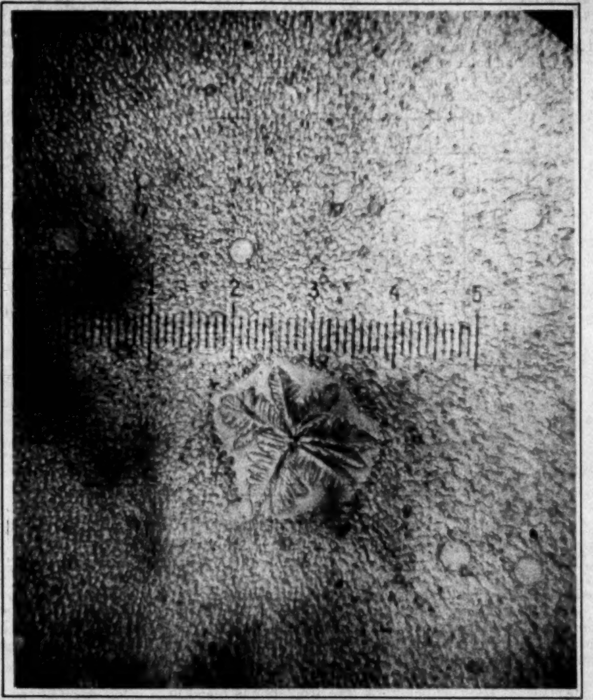

Fig. 3--Photomicrograph of formvar replicas of a snow crystal and accompanying supercooled droplets

In the investigation of the formation of snow crystals, samples of the condensates were taken at decreasing temperatures, starting with a temperature of 0°C. Different expansion ratios and expansion rates were used in order to determine whether the expansion or the temperature reached during the expansion was the important factor. The highest temperature at which snow crystals were found to form from the water vapor in ordinary air was -28.0°C. This sublimation temperature was found to be -28.0°C regardless of what expansion ratios or expansion rates were used. Figure 3 is a photomicrograph of formvar replicas of a snow crystal and a few of the accompanying supercooled droplets obtained at this temperature. This snow crystal was about 50 microns in diameter, each numbered division of the scale in Figure 3 represents 25 microns.

Table 1 correlates the temperature, at which the snow crystals were formed, with the size of snow crystal, and with size distribution and relative number of the accompanying supercooled droplets. It is seen that with lower temperatures the size of both the crystalline and liquid condensates decreases, and the number of crystals relative to that of droplets increases.

Table 1--Diameters of snow crystals and supercooled droplets at various temperatures

Temperature reached during expansion	Median diameter of snow crystal	Diameter of supercooled drops		Approximate ratio of snow crystals to drops
		Range	Median	
°C	microns	microns		
-28.0	125	4-20	9	1/200
-36.4	75	2-12	5	1/10
-38.0	70	1-3.2	2.0	1/5

It will be noted from the above table that the largest water droplet present when snow crystals first appeared (-28°C) had a diameter of 20 microns and that the smallest droplet observed (at -38°C) had a diameter of one micron. On the basis of the rate of decrease of spontaneous freezing point with decreasing size noted above, a drop of 20 microns diameter would freeze at about-38°C and a drop of one micron diameter would freeze at a temperature of the order of magnitude of -60°C. Of course such quantitative extrapolation is not warranted by these data, but it indicates qualitatively that the snow crystals observed could not have started from the spontaneous freezing of droplets in this diameter range. Such small size droplets seem to be characteristic of clouds that are produced by the cooling of moist air. SCHAEFER [1946] reported a diameter range of about five to 20 microns at about-20°C. The larger droplets that occur in swelling cumulus may be due to the continuing acquisition of water by the droplets as they fall relative to the ascending air currents.

It appears from these results that there are two ways in which the solid phase may be formed in a supercooled cloud: (1) Existing water droplets may grow large enough to freeze spontaneously. (2) The temperature of saturated air may fall to below -28°C, in which case some of the water vapor of the air will sublimate into ice crystals. To illustrate, let us assume a rising parcel of saturated air at a temperature of, say, -10°C. As the temperature of the parcel decreases, supercooled water droplets form and grow. When the temperature has reached -20°C, for instance, those droplets which have grown to a diameter of about 200 microns will freeze spontaneously (see Fig. 2). Smaller ones will remain in a supercooled state. Throughout the rise of the air parcel, new droplets are formed by condensation. When the temperature falls to -28°C, some of the water vapor will sublimate into ice crystals. At this temperature, droplets of approximately 70 microns or greater in diameter will have undergone spontaneous freezing. Further lowering of the temperature will produce relatively fewer and smaller water droplets and more ice crystals (see Table 1). In this hypothetical example, no consideration has been given to the effect of "colloidal instability" due to the coexistence of the solid and liquid phase in the same cloud.

Meteorological aspects--As to the meteorological significance of the results obtained in these investigations, the difference between the spontaneous freezing temperature and the sublimation temperature is important. As the larger supercooled cloud elements are further cooled in the rising currents, they reach the spontaneous freezing point before the rising air has reached the temperature at which its water vapor will sublimate into ice crystals. It is generally conjectured that 200 microns is the maximum diameter of droplet that can occur as the result of condensation processes alone. Assuming that the cloud droplets can grow only in this manner, the spontaneous freezing point of about -20° C of the largest cloud droplets (200 microns) thus produced would still be well above the temperature required for sublimation. Of course the growth of cloud droplets may take place by processes other than condensation such as, for example, by coalescence. It is interesting to note that a droplet of only 100 microns diameter would freeze at -25°C, a temperature that is still appreciably above the sublimation temperature. Moreover, the general temperature requirement for this freezing process agrees very well with the temperatures observed in upper portions of clouds from which active precipitation is falling, such as from cumulonimbus. FINDEISEN [1938] gives a critical temperature requirement of -15°C to -20°C for precipitating cumulonimbus.

As soon as the solid phase appears, the relatively large vapor pressure of the supercooled water droplets will cause their evaporation and crystallization onto the frozen water pellets. This crystalline sublimation onto the ice pellets was microscopically observed, and was seen to take place very rapidly. The resulting colloidal instability of the cloud initiates the precipitation. The ice crystal requirements of the Bergeron-Findeisen hypothesis are thus no longer necessary.

Certain substances, such as silver iodide, which have been used successfully in cloud-seeding experiments [VONNEGUT, 1947] raise the spontaneous freezing point of water droplets, a fact which was verified during the present investigations when a drop of approximately one mm diameter containing finely divided silver iodide was found to freeze at -4°.5 C. Thus such substances facilitate crystalline growth at higher temperatures. It also appears that cloud-seeding with dry-ice [SCHAEFER, 1946] locally chills the air to such an extent that spontaneous freezing of the supercooled droplets or even direct sublimation of the water vapor may take place.

The experimental work was performed in the laboratory of the Division of Meteorology at The Pennsylvania State College, under the direction of Hans Neuberger, Chief of the Division. The project was jointly sponsored by the Division of Meteorology and the Department of Physics.

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(Manuscript received April 13, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 22, 1948; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

THE USE OF PYRHELIOMETER RECORDS IN OBJECTIVE WEATHER FORECASTING

Charles L. Hosler

Abstract--The amount of solar and sky radiation received on a horizontal surface is used for objective forecasting of summer thunderstorms and shower type precipitation. Methods for establishing a radiation value which is critical for precipitation and ways of evaluating the records used in forecasting are explained. The results of forecasts made on this basis over a period of seven years, for the months of June, July, and August, show appreciable skill above chance.

In discussing the effect of solar radiation on precipitation, SHAW [1930] states, "... the cloudless yet nebulous condition of the atmosphere due to dust or some other nucleation that could not be called cloud is well known. ... It might well be regarded as the precursor of a thunderstorm; we are not aware of any study of the relation of summer thunderstorms to the previous condition of the atmosphere with respect to radiation."

The first part of this statement was proven valid in the course of the present investigation, in which an attempt is made to find synoptic uses of solar and sky radiation records.

Of the many factors which have an effect on the amount of radiation reaching the surface of the Earth, water in its various states causes the greatest variations. For instance, a thick overcast of a cumulonimbus cloud can cut radiation to less than one per cent of the maximum possible. On the other hand, reflection from the tops and sides of clouds increases the total radiation incident on a horizontal surface frequently by as much as 40 per cent [KIMBALL and MILLER, 1912] and occasionally may even double the radiation for a short period [IVES, 1946]. Water in the form of vapor in the atmosphere depletes radiation in proportion to the amount present. The deviation of the radiation received at the surface on cloudless days from that incident upon the outside atmosphere is related to the amount of water vapor in the atmosphere [SHAW, 1930]. In view of the response of radiation records to the moisture content of the layers penetrated, it was thought that a relation might exist between the radiation intensity recorded and the development of air-mass showers. Even though the occurrence of summer thunderstorms can be brought about by a number of processes, none of these can by themselves produce a thunderstorm or shower unless the air is sufficiently moist. In addition to the amount of water present being a factor in shower activity, the state of this water may have a great effect upon the liklihood of a shower and also upon the amount of radiation received. For instance, the growth of the condensation nuclei present in the air during the morning hours and its effect on the depletion of radiation may be an important precursor of shower activity. In this connection it must be point out, that in measuring solar and sky radiation some of the elements which deplete the direct solar radiation cause an increase in sky radiation [ÅNGSTROM, 1919], thus making the reduction less than it would otherwise be. However, the net reduction due to small droplets not recognizable as clouds may serve as an indication of the amount and availability of the moisture in the air for shower forming processes. It was desirable then, to have a measure of the integrated effect on radiation intensity of the aerosol and water in its various forms.

For this investigation, the radiation records of a conventional Eppley pyrheliometer were available. The instrument is exposed on the roof of the Mineral Industries Building at the Pennsylvania State College, at an altitude of 1200 ft above mean sea level (Lat. 40°48'N, Long. 77°52'W).

In central Pennsylvania, most convective thunderstorms and showers occur during the months of June, July, and August. These three months were therefore chosen for investigation. Inasmuch as the records were to be used for forecasting purposes, an early morning hour had to be selected. On the other hand, interference by early-morning fogs or haze had to be avoided. The hour between 0800 and 0900 TST (true solar time) was finally adopted as the key hour at which time practically all fogs had dissipated, and convective clouds were inconspicuous.

The daily maximum effectively possible solar and sky radiation was used as a standard. These values were computed for State College on the basis of the highest radiation values recorded at the station according to the method suggested by CONRAD [1929]. This method was enhanced by assuming

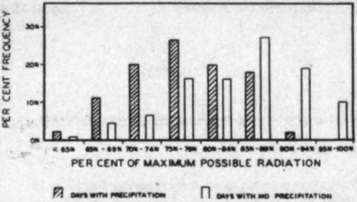

Fig. 1--Relative frequencies of radiation intensities for days with and without precipitation

symmetry about the solstices which yields a greater number of points on a graph showing the maximum radiation for every day of the year [NEUBERGER, 1945]. In order to determine the maximum radiation intensity that could occur between the hours of 0800 and 0900 TST on any particular day, the diurnal variation of the total radiation on a horizontal surface was assumed to be a sine function, symmetrical about true solar noon.

Days were investigated on which the period from 0800 to 0900 TST showed a "clear" trace (that is, undisturbed by clouds) on the pyrheliometer record. In the seven-year period from 1941 through 1947, 182 such days occurred. The number of gram calories of solar and sky radiation received during the hour from 0800 to 0900 TST on each of these days was expressed in per cent of the maximum effectively possible during this hour. The values thus obtained ranged from 59 per cent to 100 per cent. On 45 (25 per cent) of the days investigated, precipitation was recorded during the subsequent 24-hour period starting at 0900 TST. This fact in itself would indicate that if the skies are free of clouds during this key hour, the chances of precipitation are about one in four. On days on which precipitation occurred, an average value of 76 per cent of the maximum effectively possible radiation was received between 0800 and 0900 TST. On days on which no precipitation occurred, this average radiation value amounted to 84 per cent. Figure 1 shows the frequencies with which the various relative radiation intensities occurred when precipitation followed as well as when no precipitation occurred during the specified period. Both distributions are fairly normal about the mean value. In both cases 68 per cent of the values fall within one standard deviation of the mean. The overlapping of the two graphs can be explained by considering the influence of fronts and some early-morning haze and fog on radiation. More specifically, on some of the days considered here, smoke, haze, or light fog probably still persisted in densities not sufficiently great to be apparent on the pyrheliometer record, though reducing the amount of radiation reaching the instrument. On the other hand, on days when the atmosphere was not moist enough or the other requisites for condensation were not present to cause an air-mass shower, a frontal passage may have caused precipitation.

In order to utilize the above information in an attempt at objective forecasts of precipitation, a radiation value was chosen above which the chances of a shower were relatively small, and below which they were relatively great. It is apparent that any one value selected can only be approximate because of the overlapping of the two frequency distributions. The value of 80 per cent, midway between the two mean values was arbitrarily selected.

Such a criterion is obviously most useful on days when the sky is clear between 0800 and 0900 TST. However, most days are not entirely clear during this hour, so that for a more general applicability of the method, a technique must be introduced for obtaining an approximation of a clear-curve value on days when clouds interfere with radiation. Figure 2 shows two typical records of partly cloudy periods. The lowest points on the traces are instances when clouds covered the sun and the radiation received consists of sky radiation and radiation reflected from clouds. The highest peaks of the curves represent bright sunlight when the radiation incident upon the pyrheliometer disc was composed of direct solar radiation, sky radiation, and a large amount of radiation reflected from clouds [IVES, 1946; KIMBALL and MILLER, 1912]. Somewhere between these extremes is a value which represents the solar and sky radiation undisturbed by clouds. Figure 2 shows the manner in which "clear" curves were approximated for the purpose of establishing the per cent of the maximum possible radiation. In observing the response of the recorder to various sky conditions, it was noticed that with broken clouds, periods of a few minutes existed, during which the sun was not obscured and the clouds were far enough removed so that the trace assumed the character of a clear-sky record as indicated by arrows in Figure 2. Such points on the trace were used in establishing the approximate clear curves (dashed lines in Figure 2).

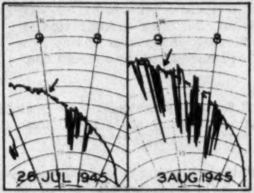

Fig. 2--Partly-cloudy radiation records and approximated clear curves

On days when the sky was completely overcast between 0800 and 0900 so that it was impossible to approximate a clear curve, the records were evaluated according to the actual amount of radiation received. Because these values were all below 50 per cent of the maximum possible, precipitation was automatically forecast. All such cases were treated separately in analysing the results of the forecasts.

By means of the method outlined above, "post-mortem forecasts" were made for every day during the summer months for the years 1941 through 1947, a total of 644 forecasts. Whenever the solar and sky radiation as evaluated from the clear curve exceeded or equaled 80 per cent of the maximum effectively possible, a forecast of no precipitation was made; when less than 80 per cent, precipitation was forecast, radiation being the only factor taken into account.

For verification of the forecasts made in this manner the following formula suggested by KÖPPEN [1906] was used. $M = (a + b - 2ab) Z$ where M is the number of misses due to chance, Z is the number of forecasts made, a is the fraction of these forecasts (Z) that called for precipitation, and b is the fraction of these forecasts (Z) on which precipitation actually occurred. Then Z - M is the number of hits due to chance.

The skill of the forecasts is then indicated by the number of hits obtained above and beyond that to be expected due to chance. The number of hits to be expected in summer at State College, Pennsylvania, happens to be very nearly 50 per cent; specifically, in this area some precipitation (including traces) occurs, on an average, during 51.5 per cent of the days of summer months. The results are shown in Tables 1, 2, and 3. Table 1 contains the analysis of forecasts made on all 644 days. The number of hits varied from as low as 62 per cent in 1945, to as high as 77 per cent in 1946. Over the seven-year period, 70 per cent hits were obtained, which is 19 per cent better than chance.

Table 1--Results of all forecasts

Year	Number of forecasts	Number of hits	Hits	Hits due to chance	Better than chance
			pct	pct	pct
1941	92	62	67	49	18
1942	92	70	76	51	25
1943	92	62	67	52	15
1944	92	60	65	50	15
1945	92	57	62	50	12
1946	92	71	77	51	26
1947	92	69	75	55	20
Total	644	451	70	51	19

Table 2--Results of clear-curve forecasts (clear and partly cloudy days)

Year	Number of forecasts	Number of hits	Hits	Hits due to chance	Better than chance
			pct	pct	pct
1941	74	48	65	54	11
1942	65	46	71	53	18
1943	75	48	64	50	14
1944	78	47	60	51	9
1945	75	41	55	52	3
1946	72	51	71	51	20
1947	67	51	76	52	24
Total	506	332	66	51	15

Table 2, which includes only days on which a clear curve could be estimated, that is, clear and partly cloudy days, shows similar results. The number of hits varied from 55 per cent in 1945 to 76 per cent in 1947. The seven-year period yielded 66 per cent hits, which is 15 per cent better than chance. Comparison with Table 1 shows that cloudiness in the early morning hours increases the precipitation probability. Precipitation occurred within 24 hours on 86 per cent of the days on which a clear trace could not be estimated and the actually recorded radiation was below 50 per cent of the maximum effectively possible. There were 138 such days.

Table 3--Results of partly cloudy forecasts

Year	Number of forecasts	Number of hits	Hits	Hits due to chance	Better than chance
			pct	pct	pct
1941	44	22	50	34	16
1942	46	31	67	49	18
1943	47	31	66	52	14
1944	45	30	67	50	17
1945	53	24	45	50	- 5
1946	39	28	72	50	22
1947	50	39	78	52	26
Total	324	205	63	49	14

Table 3 gives the results of forecasts when only the partly cloudy days are considered. These are the days on which clear curves were approximated as shown in Figure 2 and the table does not include the days used to obtain the forecasting criterion. It can be seen that with the exception of the year 1945, the results were as good or better than those obtained when the clear days were included. The data for the seven years shows that hits were obtained 14 per cent above chance. It is conceivable that even better results might be obtained by lowering the threshold value of 80 per cent as the forecasting criterion for partly cloudy days because of the effect of the clouds on the transparency of the spaces between them.

It should be emphasized that these forecasts were made on a purely objective basis without the aid of any additional meteorological data. It is not suggested that this method should be used as the sole basis for such forecasts. Nevertheless, the results demonstrate that radiation data can be used to supplement conventional methods of forecasting.

Acknowledgment--The author gratefully acknowledges the aid given in the preparation of this paper by H. Neuberger, Chief, Division of Meteorology, Pennsylvania State College.

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(Manuscript received April 13, 1948; presented at the Twenty-Ninth Annual Meeting, Washington, D. C., April 22, 1948; open for formal disucssion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

LATITUDINAL VARIATION OF MAXIMUM OBSERVED UNITED STATES RAINFALL EAST OF THE ROCKY MOUNTAINS

Robert A. McCormick

Abstract--Investigation of the latitudinal variation of observed rainfall magnitudes for specific areas and durations in over 300 major storms reveals a remarkably consistent latitudinal location of the peak magnitudes at about 30° - 31°N, abrupt declines south of the peaks and uniform variations to the north. The effect on these variations of relative opportunities for rainfall measurement and major-storm occurrence is discussed and possible dynamic causes for the observed variations are briefly outlined.

It has long been noted that the heaviest rainfalls in the United States east of the Rocky Mountains, regardless of duration, have occurred almost exclusively in the Gulf States. Of the nine storms, for instance, listed in the tabulation by SHANDS [1947] of maximum observed rainfall values in the United States, seven were centered over the Gulf States while the other two provide only four of the 97 values tabulated. However, further study of the mass of storm-rainfall data now made available by the storm-study program cooperatively pursued by the Hydrometeorological Section of the United States Weather Bureau, and the Corps of Engineers of the Department of the Army, reveals not only the expected decrease of maximum observed rainfall values from south to north but also a startling decrease southward from about latitude 30°N. These basic data include rainfall values of some 300 storms, all major and including virtually all the top storms recorded in the central and eastern portions of the country. Because precipitation from the Rocky Mountains westward is so strongly influenced by orography, values from that region were not included in this study. Nevertheless, it is interesting to note that the magnitudes of such rainfall data as are available, are in line with those indicated by the observed variations to the east.

In the analysis outlined in this paper, the maximum storm-rainfall values, stratified by area and duration, were plotted against latitude (to the nearest half degree) of the total-storm isohyetal center. The areas included in the investigation were 100, 500, 1000, 2000, 5000, and 10,000 sq mi; the durations included were 12, 24, 36, 48, and 72 hr. Figure 1 shows four examples of the plot of maximum observed rainfall against latitude, only the highest values at each latitude being shown. The numbers adjacent to the plotted points identify the storm for location on the map in Figure 2, the unnumbered points at the high latitudes, representing values from the British Isles [GLASSPOOLE, 1930], having been added as a matter of interest. A simple straight-line envelopment was made north and south of the maximum point on each chart since no adequate physical reasons have been developed which would call for a more involved type of curve.

The characteristics shown in Figure 1 are typical of those found in all of the 30 charts prepared, the plot of the values for 10,000 sq mi and 12 hr being included to illustrate the smallest relative peaking at latitudes 30° - 31°N and the greatest deviation from average shape obtained in the study. The most striking features of the charts are the consistency of the latitudinal location of the peak, the abruptness of the drop south of the peak, and the uniformity on all charts of the variation north of the peak. That the latitudinal variation in observed maximum rainfall is probably real north of the United States is indicated by the British values. On either side of latitudes for the United States, Canadian and Mexican data would have been useful in checking and extending the observed trend; unfortunately, however, those countries do not analyze their storm rainfall on an areal basis as a routine procedure.

Analysis of the latitudinal variation for areas less than 100 sq mi and durations less than 12 hr did not produce results consistent with the above. Inspection of the tabulation by SHANDS [1947] shows that for durations of 1, 6, and 12 hr at a point, and for a duration of six hr at 25 sq mi, the peak value is near latitude 42°. In weighing this apparent inconsistency, several factors should be considered: (1) Important local effects, for example, orographic, may overshadow nearly all others; (2) small-area values are the most likely to be missed by any network of gages; and (3) uncertainties in the areal representativeness of a single gage and in the time distribution of the rainfall catch of non-recording gages make the small-area and short-duration determinations the least reliable.

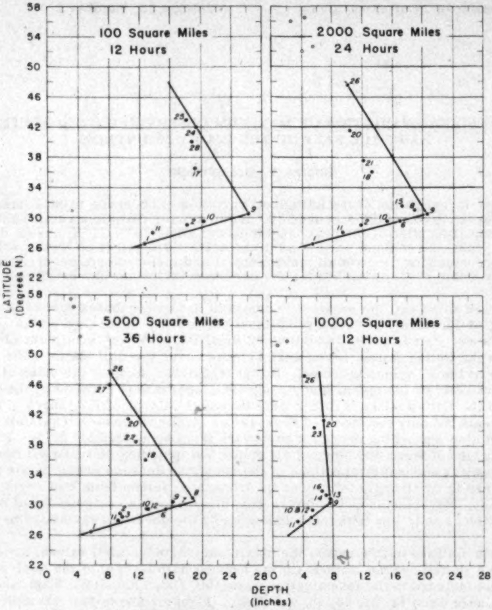
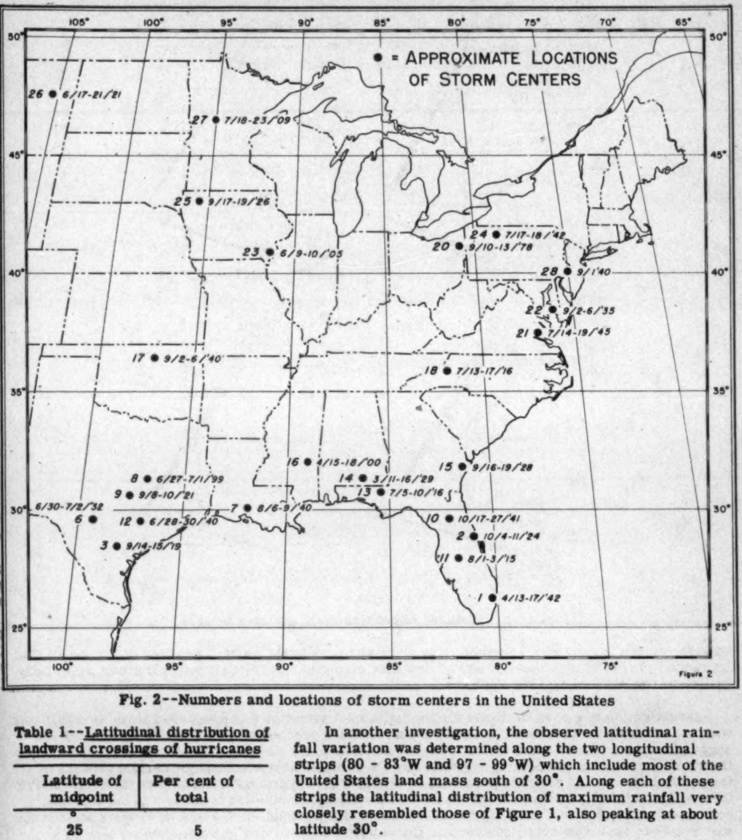


Fig. 1--Maximum rainfall observed in the United States (numbers at points identify storms, refer to Figure 2; ° = data from British Isles)

The trends of the latitudinal variation with duration and area are illustrated in Figures 3 and 4, respectively. The lines shown are envelopes of percentages of peak rainfall. For convenience of comparison, all values have been shifted to a common latitude. An exceptional regularity is found north of the peak in Figure 3, enveloping slope varying directly with duration, but the total range of slope is small. No such orderly progression is evident either by duration south of the peaks in Figure 3 or by area in Figure 4.

Even if the distribution is accepted climatologically, a legitimate question arises concerning the reliability of the observed variation as a measure of the latitudinal variation of storm-rainfall potential. Unless all the maximum observed values (or, rather, their envelopments) are equally close to the maximum rainfall possible at their respective latitudes, the latitudinal variation in potential rainfall probably differs from that shown in Figure 1. Particularly important, with respect to the southward decrease of observed rainfall from the peak at latitude 30° - 31°N, is the small amount of land area included within the United States south of those latitudes. Therefore the possibility suggests itself that the sharp observed drop is merely an effect of the lack of opportunity for rainfall measurement.

A partial evaluation of relative heavy-rainfall opportunity can be made through analysis of hurricane frequencies. Of the five storms which produced all the maximum peak values for the areas and durations considered here, two directly, and two indirectly, were associated with tropical storms. Furthermore, the greatest hurricane-rainfall intensities characteristically occur within a few miles inland from the coast. Accordingly, TANNEHILL'S [1944] tropical-storm trajectory charts were analyzed to determine the number of landward crossings within each degree of latitude south of Cape Hatteras. The distribution of the 144 crossings counted is shown in Table 1. Although the frequency peaks at latitude 30°, nearly twice as many tropical storms entered the coast between latitudes 25° and 29° as did at latitude 30°.



105°
100°
95°
90°
85°
80°
75°
70°
65°
50°
45°
40°
35°
30°
25°
● = APPROXIMATE LOCATIONS OF STORM CENTERS
26 ● 6/17-21/'21
27 ● 7/18-23/'09
25 ● 9/17-19/'26
23 ● 6/9-10/'05
24 ● 7/17-18/'42
20 ● 9/10-13/'78
28 ● 9/1'40
22 ● 9/2-6/'35
21 ● 7/14-19/'45
17 ● 9/2-6/'40
18 ● 7/13-17/'16
16 ● 4/15-18/'00
15 ● 9/16-19/'28
8 ● 6/27-7/1/'99
14 ● 3/11-16/'29
9 ● 9/8-10/'21
13 ● 7/5-10/'16
6/30-7/2/'32
6 ●
7 ● 8/6-9/'40
12 ● 6/28-30/'40
10 ● 10/17-27/'41
2 ● 10/4-11/'24
3 ● 9/14-15/'19
11 ● 8/1-3/'15
1 ● 4/13-17/'42
Figure 2

Fig. 2--Numbers and locations of storm centers in the United States

Table 1--Latitudinal distribution of landward crossings of hurricanes

Latitude of midpoint	Per cent of total
°	
25	5
26	14
27	12
28	7
29	19
30	32
31	2
32	4
33	3
34	1
35	1

In another investigation, the observed latitudinal rainfall variation was determined along the two longitudinal strips (80 - 83°W and 97 - 99°W) which include most of the United States land mass south of 30°. Along each of these strips the latitudinal distribution of maximum rainfall very closely resembled those of Figure 1, also peaking at about latitude 30°.

In maximum-storm determinations, the Hydrometeorological Section has used a technique in which a moisture factor and a wind factor are combined to produce the maximum rainfall [SHOWALTER and SOLOT, 1942]. The former is usually a precipitable-water value extrapolated from surface dewpoint and the latter a complex measure of not only the inflow-wind speed but also the wind convergence. Discussions available in recent Hydrometeorological Reports [HYDROMETEOROLOGICAL SECTION, 1946 and 1947], and in a recent paper by FLETCHER [1948] show that the relative magnitudes of the wind factors in observed major storms can be determined by dividing observed rainfall by observed moisture charge. When this is done, a latitudinal profile of the

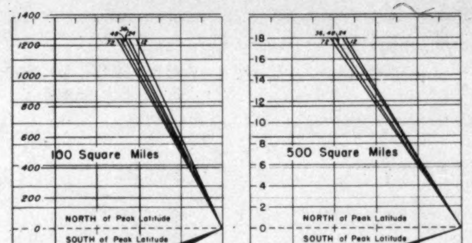


Fig. 3--Percentages of peak rainfall, grouped by area

wind factor is obtained which is similar in slope and peak to the rainfall profiles of Figure 1. The latitudinal profile of maximum observed dewpoint also peaks in the Gulf States but then apparently maintains the same high level southward.

Investigation of a physical basis for the latitudinal variation has proceeded along several lines with, so far, not much success. It is hoped that the data presented here will elicit comment and suggestions for methods of attack. A dynamic motivation for the convergence variation, and hence the rainfall variation, is possibly suggested by the theoretical discussions of certain aspects of the general circulation recently presented by ROSSBY [1947] and collaborators at the UNIVERSITY OF CHICAGO [1947]. In these discussions it is shown that if there is constancy in the vertical component of absolute vorticity, in middle latitudes there should exist a belt of relative west winds whose speeds should increase southward. Considerations involving shear instability lead to a breakdown in this distribution somewhere north of 30°, and to establishment of a wind distribution corresponding to a constant southward flux of vorticity south of the latitude of breakdown. In the limiting case, it is shown that the breakdown can occur as far south as about 30° where, because of the shape of the profile, the west-wind maximum could have its greatest strength. If the strength of the ascending motions attending the breakdown varies with the speed of the west-wind maximum, rainfall intensities, which are well correlated with the strength of vertical currents, would presumably be greatest near 30°, in agreement with the observed latitudinal distribution. Although the data used by WILLETT [1944] show the average observed west-wind maximum to be at about 31° during the Northern Hemisphere winter and at about 51° during the summer, it must be kept in mind that the peak storms in this study, most of which were summertime occurrences, were not "average" storms. The meteorological conditions under which they were produced have necessarily been the extremes of record insofar as rainfall production is concerned.

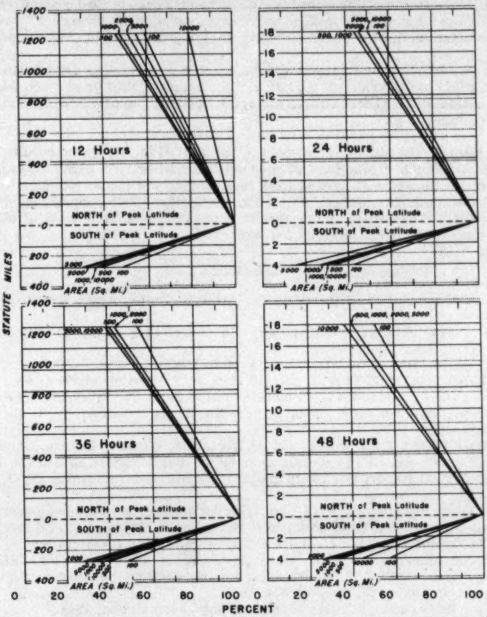


Fig. 4--Percentages of peak rainfall, grouped by duration

Qualitative correlation between the locations of the west-wind maxima and of the controlling storms in this study is almost impossible because most of the storms antedate the period of suitable upper-air observations. Although there is some evidence to indicate the presence of zonal-wind maxima above some of the more recent storm areas, investigations of these storms have not been satisfactory because of the difficulty in obtaining adequate profiles from existing networks of upper-air-observation stations.

The above discussion provides no explanation for the rapid and steady decrease of maximum observed rainfall south of the peaks. The amount of rainfall data used leaves little doubt that it is real; the remarkable similarity between the entire rainfall profile and the vorticity-theory profile suggests rather strongly that the two may be closely related. The following explanation for the observed rainfall curves is offered by A. L. Shands of the Hydrometeorological Section. The synoptic pattern predominating in maximum-storm situations can be typed as characterized by a change from anticyclonic curvature of isobars to straight or cyclonic curvature. In balanced nonfrictional flow at the same latitude, with constant pressure gradient and density, the maximum deceleration or convergence of the wind would be from the maximum anticyclonic (twice the geostrophic) to the geostrophic-wind value or, if cyclonic, to a lesser value [SHANDS, 1947]. The maximum anticyclonic wind has the value of $Rf/2$, where R is radius of curvature and f is the Coriolis parameter [HUMPHREYS, 1940]. If we now assume that the maximum value of R is potentially the same at all latitudes, or at the latitudes with which we are concerned, then the latitudinal variation of the maximum anticyclonic wind is the same as the latitudinal variation of f, which decreases from north to south. This, then, might also be the variation of the potential wind or convergence factor in maximum-storm situations. The variation of the maximum-moisture charge, on the other hand, is an increase southward to a plateau which is reached in the Gulf States. Multiplying potential convergence factor by moisture factor, at each latitude, would thus give a latitudinal profile of the

potential rainfall. And that profile, from the data at hand, bears a good qualitative resemblance to the rainfall profiles of Figure 1, with, in particular, a peak near latitude 35° and a sharp drop southward. More exact definition of the moisture-factor profile might shift the peak southward.

It is hoped that further investigations will be carried on, aimed at establishment of a more precise physical basis for the rainfall variation. The results could be of much significance with respect to explanation of the general atmospheric circulation and of particular importance in the field of hydrometeorology.

Acknowledgments--The author is indebted to the staff of the Hydrometeorological Section, particularly to R. D. Fletcher, who first pointed out the similarity between the rainfall variation and Rossby's profiles, and to A. L. Shands, for their advice and constructive criticism in the interpretation of the basic data, as well as for their assistance and encouragement in the preparation of this paper. The author also wishes to express appreciation to W. E. Kinnear for designing and directing the preparation of the figures.

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(Manuscript received May 5, 1948; presented at the Pacific Coast Regional Meeting, Berkeley, California, February 13, 1948; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

PRELIMINARY MAGNETIC CHART FOR 1947

Frank O. Klein

The accompanying chart (see Fig. 1) was prepared to indicate the general pattern of declination conditions over the area depicted, for the year 1947. It was compiled from approximately 600 declination values obtained during that year. These values were computed from information recorded by various observers flying in a number of aircraft on missions of the 72nd Reconnaissance Squadron (Very Long Range) Photographic, stationed at Ladd Air Force Base, Alaska.

The values were determined by comparing true headings obtained from an astro-compass with magnetic headings simultaneously obtained from a Fluxgate gyro-stabilized type of magnetic compass. In a large percentage of the magnetic observations, readings were averaged over a two-minute period and compared with the true heading value existing at the mid-period of the observation. Consequently, the total of individual magnetic readings considerably exceeded 600.

The average accuracy of the true heading values was on the order of plus or minus one degree. That of the magnetic values is unknown but is believed to be less than the accuracy of the true headings. The inaccuracy present is believed to be principally due to the mechanical type of compensation employed in the Fluxgate compass. A number of the compasses utilized were compensated in the vicinity of Fairbanks, Alaska, where the north component of the Earth's magnetic force approximates 0.115 CGS units. Other compasses remained free of any compensation. Theoretically, these compasses were subjected to horizontal magnetic intensities ranging from 0.115 to 0.000 CGS units. However, sensitivity studies over the charted area indicated that these theoretical values were erroneous, at least at flying altitudes.

Hunting of the compass over the area averaged an approximate value of three degrees. This was believed to be due more to aircraft oscillations than to inherent hunting of the compass.

Magnetic storms of great or very great magnitude were rarely encountered on missions flown during 1947. Magnetic observations recorded during any storm period were not used in the compilation of the chart.

Position accuracy of the individual data received thorough analyses, and errors, where found, were rectified. The resultant accuracy of positions for the individual declination values averages within five nautical miles.

Military security prevents discussion at present of the geographical distribution of magnetic observations. However, the chart was sectored into many areas, for each of which mean declination values were determined. The average probable error for the mean has not yet been calculated, although random sector computations indicate values of three and less.

The greatest inaccuracy in the chart may be expected in the vicinity of the 85th parallel and in the 90E isogonal.

This chart has been termed preliminary since the author now has additional data for its completion.

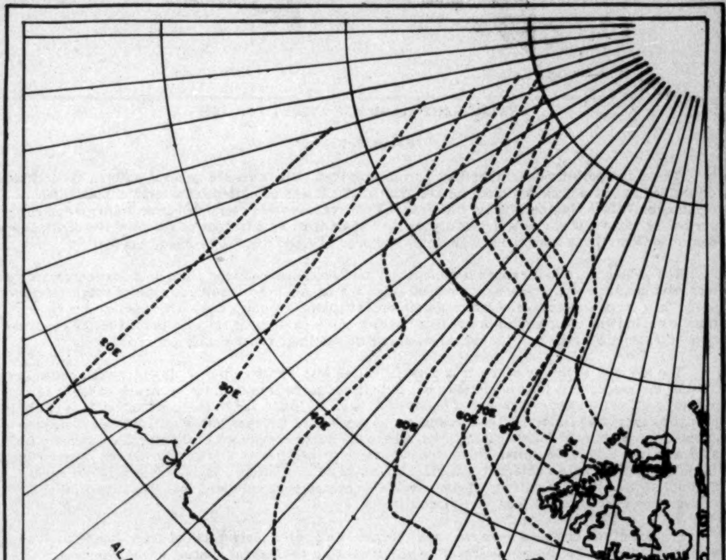
20E
30E
40E
50E
60E
70E
80E
90E
100E
ALASKA
BANKS
VICTORIA
CANADA
MELVILLE

Fig. 1--Preliminary magnetic chart for 1947 in the North American Arctic showing lines of equal declination

United States Air Force,
Fairbanks, Alaska

(Communicated manuscript received November 30, 1948; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

IMPULSIVE WAVES IN SHALLOW WATER AS GENERATED BY FALLING WEIGHTS

J. W. Johnson and K. J. Bermel

Abstract--This paper summarizes the results of observations on the characteristics of waves resulting from steel discs falling into shallow water. The variation of wave height and wave length as a function of distance from the center of disturbance is presented as a function of diameter and weight of disc and height of fall. General information is given on the efficiency of a falling weight as a wave generator.

Introduction--For planning the Bikini tests, where an atomic bomb was to be exploded in relatively shallow water, information on the characteristics of the impulsively generated surface waves resulting from such an explosion was desirable. To provide such information a series of experiments was carried out on waves generated by dropping steel discs into a body of water.

Experimental equipment and procedure--The falling-weight tests were made in the hydraulic model basin of the University of California at Berkeley. This basin (64 ft by 122 ft in plan by two feet deep) was arranged as shown in Figure 1. Weights were dropped from an overhead support and struck the water surface in the basin at point A. Radial walls, with point A as a center, were arranged to cut out a segment of the waves as they moved away from the disturbance center. Measurement of wave height was made at various distances from point A in the sector between the walls. The arrangement of walls as shown in Figure 1 permitted the tests to be conducted on the largest possible scale, that is, wave-height measurements could be made over a distance of approximately 70 ft from the center of the disturbance. Should the waves have been allowed to move unhindered in all directions for observation purposes, the tests would have been limited to a distance of only about half the width of the model basin or 32 ft from the point where the weight was dropped.

Two sets of experiments were made by dropping discs of various weights and diameters; one set (Series A) with water of uniform depth, and a second set (Series B, C, and D) with a shoaling bottom between the walls. The prototype condition assumed for these tests was that a depth of water of 160 ft would prevail at the center of the explosion. The size of the basin permitted a 1:200 scale model of the bottom of Bikini Atoll to be molded between the radial walls as shown in Figure 1. A recess in the bottom of the basin was provided so that when the weight came to rest the upper surface of the weight would be flush with the surrounding floor of the basin. This provision of a flush top gave a minimum interference with oscillation of the water after it filled the cavity "blown" into the water by the impact of the falling weight. Wave measurements were made by two methods. In the tests with a uniform depth of water only the maximum wave height above still-water level was measured, and this was accomplished by noting the wetting action on a dry stick as the waves passed. In the tests with a sector of the bottom of Bikini Island molded between the radial walls, wave heights were measured by float-type recorders, by motion pictures, and by the wetting of dry sticks as described above. The recorders were located at distances of 20, 30, and 40 ft from the center of disturbance, and the measurement of wave height by the sticks was made at ten-foot intervals over a distance of 70 ft from the center. A summary of the weights used in the tests is given in Table 1. Each weight was dropped from various heights to give a large a range in potential energy as possible. The procedure in dropping the weights was to elevate the weight to the desired height by means of a hand winch, wait until the weight stopped swinging, and then cause the weight to fall by burning through the supporting rope with a blow torch. This method was used to minimize side thrust and allow the weight to strike the water surface in a horizontal position.

Experimental results--The following symbols are used:

H = Height of wave crest above still water level, ft
R = Distance from center of disturbance to point of wave measurement, ft
d_r = Water depth at radius R, ft
d_0 = Water depth in the vicinity of the disturbance, ft
L_0 = A fictitious wave length applicable only to the first half of the first wave, ft
D_p = A maximum diameter of water column, ft
h = A height of fall of the weight, ft

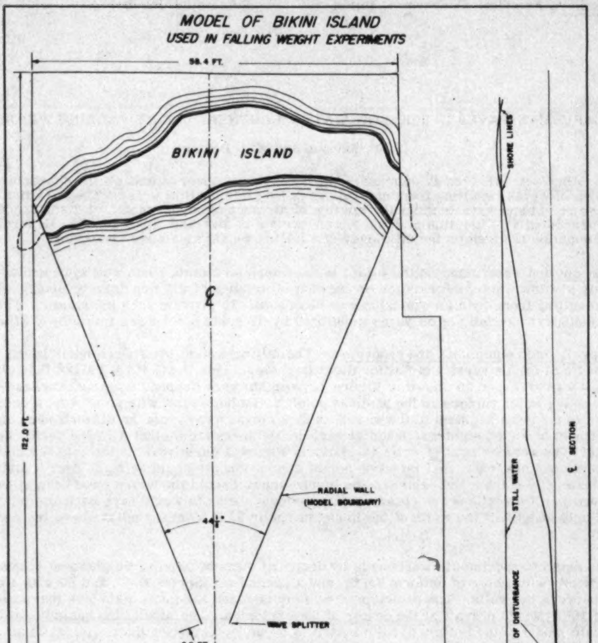
MODEL OF BIKINI ISLAND
USED IN FALLING WEIGHT EXPERIMENTS
98.4 FT.
122.0 FT.
BIKINI ISLAND
SHORE LINES
℄
RADIAL WALL
(MODEL BOUNDARY)
44½°
℄ SECTION
STILL WATER
WAVE SPLITTER
6.0 FT.
A
CENTER OF DISTURBANCE
(DOES NOT CORRESPOND TO POSITION OF FULL SCALE TEST)
BOUNDARY - MODEL BASIN
CENTER OF DISTURBANCE
0.8 FT. IN MODEL
160 FT. IN PROTOTYPE
℄
64.4 FT.
SCALES IN FEET
MODEL 0 5 10 20 30
PROTOTYPE 0 2000 4000 5880 6000

Fig. 1--Basin for a study of impulsive waves generated by weights falling in a model of Bikini Lagoon

A summary of the test conditions for each weight used in the investigation appears in Table 1. As shown in this table, discs of various diameters, weights, and bottom shape were used. The data observed in these tests are rather voluminous and only typical examples are discussed below to illustrate the more important conclusions. Those tests summarized herein were selected on the basis that the ratio, D_p/d_0, bracketed the value of this ratio as stated in newspaper accounts of test Baker at Bikini (that is $D_p/d_0 = 11$, approximately). Tests in this categroy are summarized in Table 2.

Table 1--Summary of conditions in falling weights experiments

Series and run	Data on disc				Height of fall	Potential energy	Remarks
	Weight of disc	Diameter	Center thickness	Bottom shape of disc			
	lb	ft	in		ft	ft lb	
A-1	132	2.25	13/16	Flat	3.67	484	Uniform depth, 1.02 ft
2	132	2.25	13/16	Flat	7.03	924	
3	132	2.25	13/16	Flat	14.5	1913	
4	368	4	13/16	Flat	5.4	1990	
5	368	4	13/16	Flat	8.1	2980	
6	368	4	13/16	Flat	10.9	4010	
7	132	2.25	13/16	Flat	15.4	2030	
8	132	2.25	13/16	Flat	3.8	502	
9	132	2.25	13/16	Flat	7.6	1003	
10	132	2.25	13/16	Flat	15.4	2030	
11	132	2.25	13/16	Flat	15.4	2030	Uniform depth, 0.8 ft
12	368	4	13/16	Flat	11.0	4048	
13	368	4	13/16	Flat	8.1	2980	
14	368	4	13/16	Flat	5.4	1990	
15	132	2.25	13/16	Flat	7.6	1003	
16	132	2.25	13/16	Flat	3.8	502	
B-1	1014	4	4 5/16	Curved	4.7	4770	Sloping bottom, depth 0.8 ft at hole
2	1014	4	4 5/16	Curved	9.6	9730	
3	1014	4	4 5/16	Curved	14.3	14500	
4	1014	4	4 5/16	Curved	2.7	2740	
5	1014	4	4 5/16	Curved	15.5	15700	
6	1014	4	4 5/16	Curved	15.5	15700	
7	1014	4	4 5/16	Curved	0.75	760	
8	1014	4	4 5/16	Curved	11.6	11700	
9	1028	4	4 5/16	Curved	3.7	3800	
10	1028	4	4 5/16	Curved	3.7	3800	
11	1028	4	4 5/16	Curved	9.5	9760	
12	1028	4	4 5/16	Curved	15.3	15730	
B-13	1028	4	4 5/16	Curved	1.7	1750	
14	1028	4	4 5/16	Curved	5.6	5760	
15	156	2.25	4 5/16	Curved	6.4	1000	
16	156	2.25	4 5/16	Curved	12.8	2000	
17	156	2.25	4 5/16	Curved	16.0	2500	
18	156	2.25	4 5/16	Curved	1.9	295	
19	156	2.25	4 5/16	Flat	6.4	1000	
20	156	2.25	4 5/16	Flat	12.8	2000	
21	1041	4	4 5/16	Flat	0.85	885	
22	1041	4	4 5/16	Flat	1.8	1875	
23	1041	4	4 5/16	Flat	3.8	3960	
24	1041	4	4 5/16	Flat	5.7	5930	
25	1041	4	4 5/16	Flat	9.6	10000	
26	1041	4	4 5/16	Flat	15.5	16130	
27	135	2.25	4 5/16	Flat	2.2	297	
28	135	2.25	4 5/16	Flat	7.4	1000	
C-1	375	4	4 5/16	Flat	2.7	1013	
2	375	4	4 5/16	Flat	5.4	2025	
3	375	4	4 5/16	Flat	10.8	4050	
4	375	4	4 5/16	Flat	16.2	6075	
5	375	4	4 5/16	Flat	2.7	1013	

(Continued)

Table 1--Summary of conditions in falling weights experiments (Concluded)

Series and run	Data on disc				Height of fall	Potential energy	Remarks
	Weight of disc	Diameter	Center thickness	Bottom shape of disc			
	lb	ft	in		ft	ft lb	
6	375	4	4 5/16	Flat	5.4	2025	Sloping bottom, depth 0.8 ft at hole
7	375	4	4 5/16	Flat	10.8	4050	
8	375	4	4 5/16	Flat	16.2	6075	
D-1	375	4	4 5/16	Flat	2.7	1013	Coral mound at R = 10 ft
2	375	4	4 5/16	Flat	5.4	2025	
3	375	4	4 5/16	Flat	10.8	4050	
4	375	4	4 5/16	Flat	16.2	6075	

Table 2--Experiments in which the ratio, D_p/d_0, approximates that for Test Baker

Run	D_p	d_0	D_p/d_0
	ft	ft	
A-12	7.90	0.8	9.9
A-14	6.66	0.8	8.3
B-21	5.82	0.8	7.8
B-22	6.32	0.8	7.9
B-23	7.44	0.8	9.3
B-25	10.82	0.8	13.5
B-26	12.10	0.8	15.2

It is of interest to note that weights with a flat bottom surface gave a water column with almost vertical sides as shown in Figure 2, this appearance being similar to actual explosions in shallow water; whereas a weight with with a curved bottom surface forced the water outward at a considerable angle and, therefore, was not considered as representative of an explosion.

The various runs listed in Table 2 pertain to four-foot diameter discs weighing 368 lb and 1041 lb, both of which had flat bottom surfaces. A summary of the wave heights, H, observed at various distances, R, from the center

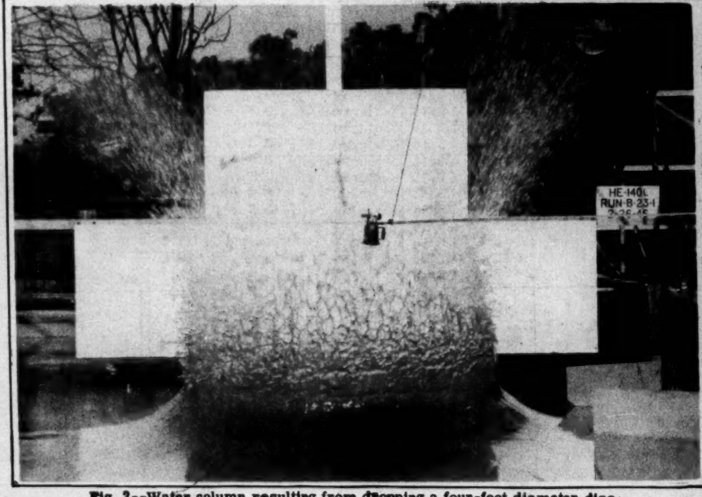

Fig. 2--Water column resulting from dropping a four-foot diameter disc into water 0.8 feet deep (Run B-23)

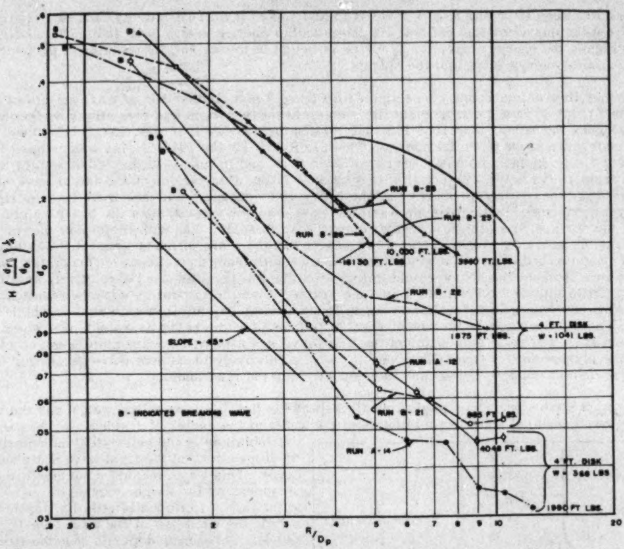

Fig. 3--Variation in wave height with distance from center of disturbance, waves generated by four-foot diameter discs falling various heights

of the disturbance is given in Figure 3 for these tests. Because the data for Runs B-21, B-22, B-23, B-25, and B-26 pertain to observations on a sloping beach, the wave heights observed at various radii have been corrected to give wave heights in a depth, d_0. The correction factor in this instance was arbitrarily assumed as the fourth root of the ratio of the depth at the point of wave measurement to the depth in the vicinity of the disturbance. To give a dimensionless representation of the variation of wave height with distance, the ratio of the corrected wave height to the depth at the disturbance is plotted in Figure 3 against the ratio of the radius at the point of wave measurement to the diameter of the water column, that is, R/D_p. The wave height in all instances was the maximum height of crest above still water.

Figure 3 is of interest because of the nature of the variation of wave height with distance from the center of the disturbance, and because of the information on the efficiency of falling weight as wave generators. With respect to wave heights it is noted that the height appears to vary with distance according to one law for values of R/D_p less than about five, and then follows another law for greater values of R/D_p. This break in the curves of height versus radius occurs for the tests with a uniform depth as well as with a shoaling bottom. It is to be noted that the first observation of wave height was at ten feet from the center of disturbance, and the wave always was breaking at this point; hence, the initial point for each set of data is indicated as a breaker on Figure 3. Because there is considerable scatter in the data, no attempt has been made to obtain an equation for the relationship between wave height and radius; hence, a broken line has been drawn to connect all points merely to give information on general trends.

The various curves in Figure 3 are of value in giving qualitative information on the efficiency of falling weights as wave generators. The curves for Runs B-21 to B-26, inclusive, resulted from dropping a four-foot diameter flat disc, weighing 1041 lb, from various heights. The potential energy of the plate, prior to release, is indicated on each curve. Examinatior of Figure 3 shows that the wave height at a particular radius increases as the potential energy is increased until a certain value of potential energy is reached, beyond which little increase in wave height occurs. Thus, the waves for Runs B-26 and B-25 (10,000 ft lb and 16,130 ft lb, respectively) are not greatly different from the curve for Run B-23 where the potential energy is only 3960 ft lb. This indicates that the higher the energy content is by virtue of height of the weight, the greater is the percentage of the available energy going into the splash.

Another item of interest in connection with Figure 3 is a comparison of wave heights occurring at the first point of wave measurement (R ten feet, where bottom was essentially horizontal in all tests) when the lighter four-foot disc (368 lb) and the heavier four-foot disc (1041 lb) were given approximately the same potential energy. Thus, in Run B-23 the 1041-lb disc was dropped from a height of 3.8 ft, giving a potential energy of 3960 ft lb, and in Run A-12 the 368-lb weight was dropped from 11 ft to give a potential energy of 4048 ft lb. Comparison of the initial wave heights in Figure 3 for these two runs shows that the heavy disc generated a higher wave than the light disc. In this instance the light weight was moving at a higher velocity than the heavy weight upon striking the surface, thus giving a greater energy loss in splash than with the slower moving heavy disc. That the heavy disc falling from a low height was the more efficient generator was evident also in connection with the secondary plume which results from the collapse of the crater formed in the water. Because the heavy weight falling from the low height caused a relatively small splash, there was little interference with flow into the crater formed in the water, with the result that the secondary plume was well defined and rose to much greater heights than with runs in which the potential energy was large by virtue of height of fall. It is of interest also to note that when the wave generation was relatively efficient the first wave was always the maximum in height and remained the highest until it reached the beach. With a relatively inefficient wave generation (high velocity of disc at impact), the second wave usually was the maximum.

The first wave always was pushed out directly by the first impulse of the weight and the waves which followed resulted from the oscillation of the water at the center of disturbance. Some of the data obtained in the tests yield information on the dimensions of the first wave in its initial stage. This wave, pushed up to maximum steepness by the weight moving through the water, has a fictitious length, L_0, applicable only to the first half of the wave (see Fig. 2 and 4). This length depends upon the time required for the weight to push through the water; that is, L_0 probably is a function of the diameter of the column of water, D_p; the depth of water, d_0; and the velocity of the weight at impact (the fall velocity being proportional to the height of fall, h; that is, $V = \sqrt{2gh}$ in a vacuum). Dimensional considerations indicate that these forms could be related as follows,

$$D_p/L_0 = f\ (h/d_0)$$

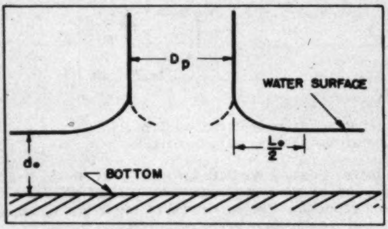

Fig. 4--Sources of the dimensions used in Table 3 and Figure 5

For the B runs listed in Table 2 it was possible to measure D_p and L_0 from photographs. These values along with the values of d_0 and h are given in Table 3 along with values of D_p/L_0 and h/d_0. A plot of these latter terms appears in Figure 5. Although the data are rather limited, this plot shows the general order of magnitude of the variables for four-foot diameter plates.

Table 3--Initial length of first wave

Run	D_p	d_0	$L_0/2$	h	D_p/L_0	h/d_0
	ft	ft	ft	ft		
B-21	5.82	0.8	1.04	0.85	2.8	1.1
B-22	6.32	0.8	1.09	1.8	2.9	2.3
B-23	7.44	0.8	1.27	3.8	2.9	4.8
B-25	10.82	0.8	1.64	9.6	3.3	12.0
B-26	12.10	0.8	1.88	15.5	3.2	19.4

In addition to the results discussed above, the falling weight tests provide data on the velocity of the waves as they moved away from the disturbance center. A typical set of data is shown in Figure 6 where the position of the first three waves is plotted against the time after the disc struck the water surface. The slope of these curves at any point gives the wave velocity at that point. The wave length at any time after explosion is represented in Figure 6 by the vertical distance between any two adjacent curves. More complete data on the variation in wave length and wave velocity with time after disturbance than available in the experiments described above are the results from a series of investigations on impulsively generated waves made in the same model basin on a previous occasion by R. G. Folsom. In these tests impulsively generated waves were produced in different constant depths of water by dropping cylinders in a vertical position from a height of about 1 1/2 ft above the water surface. The test conditions are shown in Table 4.

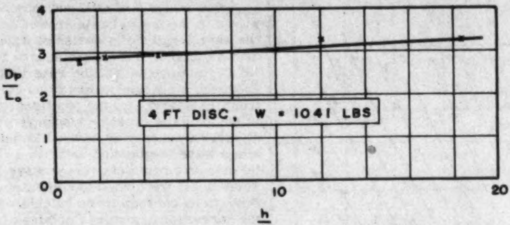

Fig. 5--Relationship between diameter of water column, length of first wave, and initial height of disc

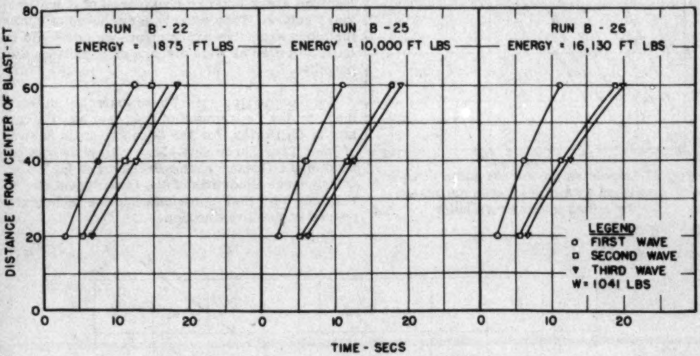

Fig. 6--Variation of wave position with time, waves generated by a four-foot disc dropped from various heights

Table 4--Test conditions for Folsom's experiments

Water depth ft	Cylinder diameter ft
0.15	0.55
0.25	0.55
0.46	0.95

The cylinders were dropped at the approximate center of a 30 ft by 38 ft clear area in the center of the model basin. The resulting phenomena were recorded on 16 mm movie film and with a still camera. The surface shadow of a rope equipped with one-foot markers was used to indicate the crest and trough positions of the resulting waves. The cylinders were dropped in a position to intersect the surface shadow of the rope. Due to the surface disturbances, distortions in the linear scales of the film occurred. As a result, no measurements of wave heights were attempted and distance from the center of impact was recorded as the length from center to the apparent wave crest using an average scale based on the overall indicated length in the photograph. The time in the moving pictures was obtained from photographs of an electric clock with a high speed hand. A typical example of the results of the measurements made from the film is shown in Figure 7. The lines in this graph are drawn to represent the successive positions of the waves at various times after the start of the generation of the initial wave. Other information appears in Figure 8 which shows wave lengths determined from crest to crest or trough to trough plotted against the distance from the center.

For these experiments, the source covered a definite area, the energy available was limited, and the generating center was a solid plunger which remained in the hole blown in the water. The impulsive waves generated consisted of an initial wave of maximum amplitude followed by a long trough and then a series of crests decreasing in length as one proceeded back toward the source.

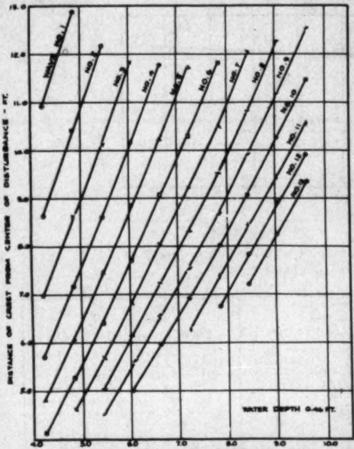

Fig. 7--Impulsive waves, variation of wave positions with time, waves generated by a 0.95 diameter cylinder

The analysis of still and moving pictures of waves in the model basin showed (a) at a given time, the wave lengths of a series of waves vary directly as the distance X from the source (see Fig. 8), (b) as time increases, the wave lengths tend to approach a constant value at the larger distances from the source, (c) the positions of given waves (distance X of crest or trough from the source) at various times may be used to determine average wave lengths and velocities (see Fig. 7), (d) measurements of average wave length and water depth were used to compute wave velocities. Deviations between these calculated values and the corresponding values obtained from the slope of distance-time curves (see Fig. 7) were small.

Conclusion--The various experiments summarized above must be considered of a preliminary nature. They merely serve as pilot studies for more extensive experiments which should be made to cover as wide a range of conditions as possible.

Acknowledgments--These experiments were made by the Department of Engineering, University of California, for the Oceanographic Section of Joint Task Force One. Roger Revelle was head of the Oceanographic Section and M. P. O'Brien was Chairman of the Department of Engineering. Both contributed to the technical phases of the investigation.

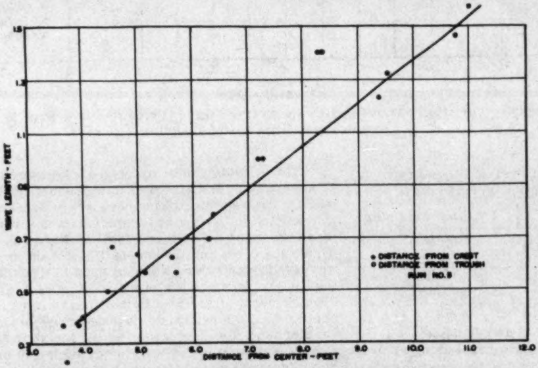

Fig. 8--Variation of wave length with distance from center of disturbance for impulsive waves; time after impulse, 5 3/4 sec estimated

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(Manuscript received June 15, 1948; presented at the Twenty-Eighth Annual Meeting, Washington, D. C., April 28, 1947; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

THE USE OF BATHYTHERMOGRAMS TO DETERMINE OCEAN CURRENTS

E. C. LaFond

Abstract--Bathythermograph observations taken during Operation Crossroads were studied to determine the accuracy and limitations of a method deriving ocean currents from these data. The vertical density distribution, necessary to determine relative currents, was obtained from bathythermograms by relating temperature to salinity.

Introduction--The conventional method of determining relative currents in the ocean was established by BJERKNES [1910], and by SANDSTROM and HELLAND-HANSEN [1903]. The calculations were simplified by SVERDRUP [1933]. This method requires a vertical series of temperature and salinity observations, obtained by means of reversing thermometers and Nansen bottles. During these observations the ship must be lying to, for as many as two to four hours.

In this paper the use of bathythermograms is proposed to determine relative currents. To use bathythermograms alone a temperature-salinity (T-S) relation must be established, which may be obtained from existing hydrographic data or be established by sampling. The accuracy of relative currents determined from temperature data would therefore be directly dependent on the reliability of the T-S relation. Expense, manpower, and time could be saved if ocean currents could be obtained from the vertical temperature structure. These savings must be weighed against the decreased accuracy of results obtained by this method.

An opportunity to investigate the use of bathythermograms in determining relative currents was provided by the Crossroads oceanographic program. At Operation Crossroads a large number of bathythermograph lowerings were made in addition to the vertical series of temperature and salinity observations. These lowerings were made for several purposes. A 900-ft bathythermograph lowering was made at each hydrographic station to check the reversing thermometer reading. Other bathythermograph observations were made to measure the detailed thermal structure in the upper layers around the Marshall Islands. Repeated observations at one of the stations were made to determine the vertical fluctuations in temperature.

Bathythermograms were also used in an attempt to establish the direction of surface current flow near the Marshall Islands. The following discussion deals principally with this latter use of bathythermograms in low latitudes.

Relative currents--From temperature-salinity-depth measurements relative currents determined by the conventional method are computed from the vertical temperature and salinity structure of the water. The derivation of the formula is based on the assumptions that (1) the vertical acceleration of individual water particles is zero, (2) the frictional term is relatively small and can be neglected, and (3) the hydrostatic equation sufficiently defines the vertical force field. With these assumptions the equation of horizontal motion becomes

$$v = -(\alpha \partial p/\partial x)/2\omega \sin \phi \quad \text{(1)}$$

where α is the specific volume of the water, $\partial p/\partial x$ is the pressure differential with respect to a horizontal axis, ω is the angular velocity of the Earth's rotation and ϕ is the latitude. By employing the hydrostatic equation, the pressure gradient may be replaced by the geopotential slope of the isobaric surfaces. If A and B represent two locations between which the current is to be computed (1) becomes

$$v_{1,2} = 10\,(D_A - D_B)/L2\omega \sin \phi \quad \text{(2)}$$

In this equation L is the distance between the two locations, D_A and D_B are the geopotential heights of the surface over some reference pressure level, and $v_{1,2}$ is the velocity at the surface with respect to the velocity at the depth of the reference level. A reference level should be chosen at a depth where the current is negligible, so the $v_{1,2}$ represents the total surface current with respect to the mass field.

The geopotential distance, D, is given by

$$D = \int_{P_1}^{P_2} \alpha_{s,\vartheta,p}\, dp \qquad (3)$$

where $\alpha_{s,\vartheta,p}$ is the specific volume of the water and is a function of salinity, temperature, and pressure. In practice the specific volume is referred to a standard specific volume, usually the value at 35 °/oo salinity, 0°C temperature, and pressure p. The specific volume is then

$$\alpha_{s,\vartheta,p} = \alpha_{35,0,p} + \delta \qquad (4)$$

where δ is the specific volume anomaly. Since the surfaces of $\alpha_{35,0,p}$ are parallel to the geopotential surfaces, the only contribution to a difference in geopotential height between two stations will be from the specific volume anomaly.

If

$$\Delta D = \int_{P_1}^{P_2} \delta dp \qquad (5)$$

is defined as the geopotential anomaly, then

$$D_A - D_B = \Delta D_A - \Delta D_B \qquad (6)$$

and the equation for the speed of the current becomes

$$v = 10\,(\Delta D_A - \Delta D_B)/L2\omega\sin\phi \qquad (7)$$

Temperature-salinity relation--To use bathythermograms to determine relative currents in the ocean, the vertical mass field must be established from the vertical temperature distribution. This can be accomplished by relating salinity to temperature. With the salinity as a function of temperature, the specific volume can be expressed as a function of temperature and pressure alone.

$$\alpha = \alpha(s, \vartheta, p) = \alpha(s(\vartheta), \vartheta, p) = \alpha(\vartheta, p) \qquad (8)$$

The geopotential distance, D, can then be expressed as a function of temperature and pressure alone.

To establish the particular T-S relation for the Bikini area in summer, salinity was plotted versus corresponding reversing thermometer temperatures (see Fig. 1). This graph was a composite of data from 107 Crossroads hydrographic stations between 10° and 13°N in the vicinity of the Marshall Islands. At each station 13 simultaneous temperature and salinity observations were normally made. From this composite of over 1000 simultaneous values of temperature and salinity, a mean T-S relation was determined by averaging the salinity observations by one degree temperature intervals. Above 27°C, half degree intervals were used to get the average salinity. This relation is represented by the average curve in Figure 1. An average salinity for any temperature can be read from this curve.

The standard error of estimate about the average temperature-salinity curve is 0.10 °/oo. The greatest spread of salinity from the mean T-S curve occurs at the highest temperature, which is near the surface (see Fig. 1). The uncertainty of the surface salinity can be overcome by collecting a surface water sample. By titrating the water sample for salinity, a definite salinity fix can be established at the surface. This salinity fix can be extrapolated accurately through the isothermal layer (see Fig. 2). The least deviation from the mean T-S curve occurs between 13° and 24°C. Below 10°C the deviation is greater. These temperatures, however, are below the range of the 900-ft bathythermograph and are not used in determining the relative currents from bathythermograms.

If the average salinity is in error by 0.10 °/oo, the calculated value of the dynamic height anomaly (0/305 dynamic meters) would be in error by 0.0022 dynamic meters. If the salinities at all depths were in error in the same direction by 0.10 °/oo, the corresponding error in dynamic height would be 0.02 dynamic meters. This latter assumption is extremely unlikely; nevertheless, the resultant error would be less than two per cent of the total calculated dynamic height values (0/305 dynamic meters) in this area.

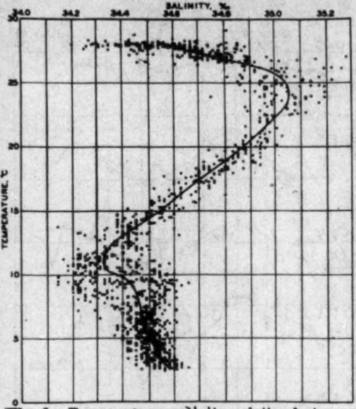

Fig. 1--Temperature-salinity relation between 10° and 13°N in the vicinity of the Marshall Islands

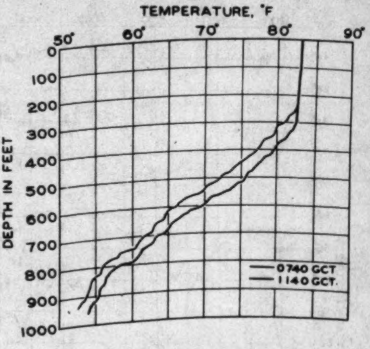

Fig. 2--Two superimposed bathythermograms, taken at 11°N 166°E, showing the change in vertical temperature structure that occurred in four hours

Certain variations in the T-S relation can be related to the geographic position. BARNES, BUMPUS, and LYMAN [1948] have pointed out the latitude variations in the T-S relation in the region near Bikini. Similar studies off the coast of Southern California and near Georges Bank demonstrated that the T-S relation varies appreciably with geographic position. In the limited area under consideration, however, this change in the T-S relation was too small to take into account. Variations in the T-S function at any one latitude were greater than the difference between observations taken at the extreme latitudes (10° to 13°N).

In high latitudes a modification must be made to the T-S relation for seasonal effects. Seasonal radiation, evaporation, and other meteorological phenomena as they affect ocean temperature and salinity are relatively uniform in low latitudes, however. Temperatures in the surface layer are virtually constant; consequently no modification for seasonal effect was made to the average T-S curve.

Internal waves--The greatest factor in producing variations in the observed density structure of the water column, thus affecting current determinations, is the vertical fluctuation of temperature and salinity commonly known as internal waves. There is a large transitional layer separating the relatively light surface layer water from the heavier deeper water. An internal wave causing a change in the depth of this transitional layer materially changes the vertical mass field. Internal waves are therefore important factors in the determination of relative currents.

Internal waves are not wholly random, but appear to fall in a cyclic pattern. This general trend is shown by dotted lines on the temperature curves (see Fig. 3). The high phases fall about 12 hr apart and have nearly the same period as the tide. The greatest changes in temperature, as well as salinity, occur around 700 ft below the surface. Above a depth of 400 ft the changes are small. At 900 ft below the surface, changes in temperature indicative of vertical fluctuations are somewhat smaller than those at 700 ft. The decrease in the magnitude of temperature variations below a depth of 700 ft indicates that the effect of internal waves on the vertical mass field is greatest at depths less than 700 ft.

Calculation of currents--After consideration of all the factors discussed, the calculation of currents from bathythermograms was attempted. By use of the temperature-salinity relation the specific volume anomaly δ was converted from a function of temperature, salinity, and depth to a function of temperature and depth only. The numerical values of $10^5\delta$ over the temperature-depth range encountered are presented in Table 1.

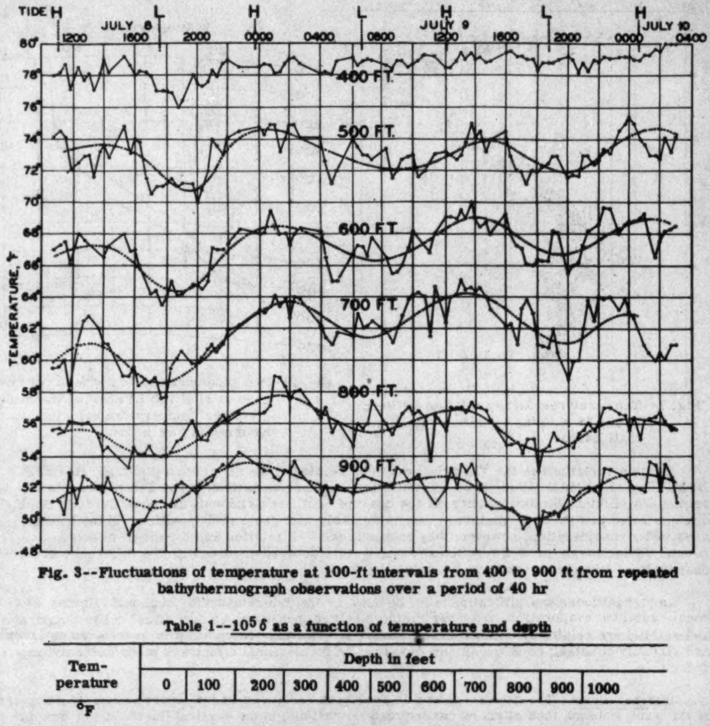

Fig. 3--Fluctuations of temperature at 100-ft intervals from 400 to 900 ft from repeated bathythermograph observations over a period of 40 hr

Table 1--$10^5 \delta$ as a function of temperature and depth

Temperature	Depth in feet										
	0	100	200	300	400	500	600	700	800	900	1000
°F											
82	565	566	567	568	570	...	...	...	...	...	...
80	514	515	516	517	519	520	...	...	...	...	...
78	469	470	471	472	474	475	...	...	...	...	...
76	...	...	...	...	438	439	440	...	...	...	...
74	...	...	...	...	...	407	409	...	...	...	...
72	...	...	...	...	...	381	382	384	...	...	...
70	...	...	...	...	...	356	357	359	...	...	...
68	...	...	...	...	...	332	333	334	...	...	...
66	...	...	...	...	...	310	311	312	...	...	...
64	...	...	...	...	...	...	289	290	291	...	...
62	...	...	...	...	...	...	272	273	274	...	...
60	...	...	...	...	...	...	...	251	252	253	...
58	...	...	...	...	...	...	...	234	235	235	236
56	...	...	...	...	...	...	...	217	218	219	219
54	...	...	...	...	...	...	...	201	202	203	203
52	...	...	...	...	...	...	...	183	184	184	185
50	...	...	...	...	...	...	...	160	160	161	162
48	...	...	...	...	...	...	...	138	138	139	139

Temperatures were read at arbitrarily chosen 100-ft intervals between zero and 1000 ft. Since a 900-ft bathythermograph records only to 950 ft, the temperature traces were extrapolated to 1000 ft. With the temperature at each 100-ft interval of depth, a corresponding value of $10^5 \delta$ was read directly from the table. By simple numerical integration over the depth of the water column, the dynamic-height anomaly, $10^5 \Delta D$ was found by means of (5). Flow is parallel to contours of D with a speed given by (7).

Discussion of results--The dynamic-height anomalies were plotted and contoured (see Fig. 4). They show a complicated pattern of flow. The general trend of flow, however, is to the west in the northern and southern part of the area, with a southern component east of the atolls. The irregular pattern can be attributed largely to internal waves.

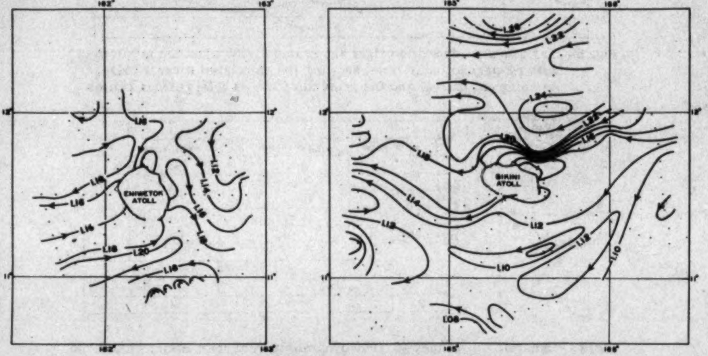

Fig. 4--Dynamic topography indicating current flow near Eniwetok and Bikini Atolls based on individual bathythermograms

To show the effect of internal waves upon current calculations, the dynamic-height anomalies were obtained from one day of repeated bathythermograph observations. The data were averaged and plotted (see Fig. 5). The resulting fluctuations in dynamic height anomalies amount to as much as 0.08 dynamic meters in a few hours.

Before considering a means of eliminating the internal wave effect, the apparent periodicity of internal waves should be mentioned. To demonstrate this periodicity the calculated dynamic topography shown in Figure 5 was subjected to a Fourier harmonic analysis [WHITTAKER and ROBINSON, 1924]. The most obvious harmonic constants of 12 and 24 lunar hours were derived. For a 24-(hour) term series, the resulting expression for dynamic height anomaly was found to be

$$\Delta D = a_0 + a_1 \cos \theta + a_2 \cos 2\theta + \ldots + a_{12} \cos 12\theta$$

$$+ b_1 \sin \theta + b_2 \sin 2\theta + \ldots + b_{11} \sin 11\theta \quad \ldots\ldots\ldots\ldots (9)$$

where ΔD is the dynamic height anomaly; $a_0, a_1, \ldots a_{12}, b_1, \ldots b_{11}$ are the coefficient for 0, 1, . . ., 12 cycles in the lunar day; and θ is the phase angle after high tide (15° = one lunar hour). This series may be expressed as

$$\Delta D = c_0 + c_1 \cos(\theta - \beta_1) + c_2 \cos(2\theta - \beta_2) + \ldots + c_{12} \cos(12\theta - \beta_{12}) \quad \ldots\ldots (10)$$

where $c = \sqrt{a^2 + b^2}$ and $\beta = \cot^{-1} a/b$. By substituting the calculated values for c and β, the individual diurnal and semidiurnal components,

$$\Delta D_{24} = 1.234 + 0.0226 \cos(\theta - 256^\circ) \quad \ldots\ldots\ldots\ldots\ldots\ldots (11)$$

and

$$\Delta D_{12} = 1.234 + 0.0186 \cos(2\theta - 21^\circ) \quad \ldots\ldots\ldots\ldots\ldots\ldots (12)$$

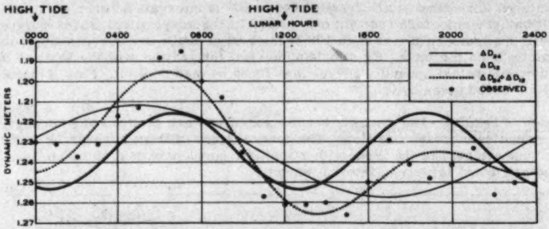

Fig. 5--Variations of dynamic-height anomalies (0/305 dynamic m) plotted with respect to lunar time, showing the calculated diurnal (ΔD_{24}), semidiurnal (ΔD_{12}) and the resultant ($\Delta D_{24} + \Delta D_{12}$) lunar cycles

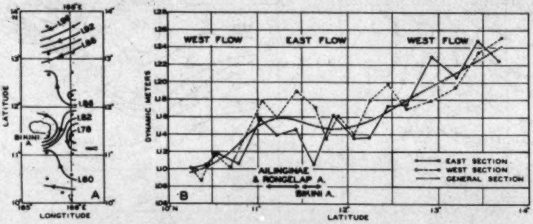

Fig. 6--Dynamic topography (0/1000 dynamic meter) near Bikini Atoll from hydrographic data (A) compared with sections of dynamic height anomalies (0/305 dynamic meter) from bathythermograms in the same area (B)

were evaluated and plotted. The sum of the diurnal and semidiurnal components were compared to the original values in Figure 5. This comparison showed that the semidiurnal component of dynamic height was a minimum about 6.5 hr after surface high tide.

Similar internal wave studies made from temperature data collected at 14 other stations in the Pacific and Atlantic oceans also show that internal waves usually have a semidiurnal period with an internal high a few hours after the surface high tide. If the amplitude and phase of the internal waves can be calculated at the time of a bathythermogram-current survey, compensations can be made for their effects on calculated currents.

A different correction for internal waves was employed for the Crossroads data. To eliminate the internal wave effect, two dynamic topography sections from bathythermograms were superimposed (see Fig. 6B). Fluctuations in the topography are apparent. By constructing a smooth curve, through these sections, however, the internal-wave effect was partially eliminated. The resultant curve shows a dip to the south below the Atoll and again between 12°.5 and 14° N. These dips are indicative of a west-flowing current. Just above the Atoll the dip is slightly to the north, indicative of an eastward flow. In this region the dynamic heights of the east section are lower than those in the west section, creating a slope to the west. This slope indicates a flow to the south just east of Bikini. These current directions are in excellent agreement with those derived from the temperature and salinity data (see Fig. 6A). The total difference in dynamic height between the north and south ends of the section was 0.19 dyn m (0/1000 dynamic meter) from the hydrographic data and 0.14 dyn m (0/305 dynamic meter) from the bathythermograph data. This corresponds to an average westerly flow of 0.33 and 0.25 knots respectively. In other words the speed of surface currents derived from bathythermograms is about 75 per cent of the speed derived from the hydrographic data.

Conclusions--To use bathythermograms in determining relative currents, there are several prerequisites which must be met: (1) The bathythermogram must extend at least 900 ft; (2) temperature-salinity relation must be established, with consideration of seasonal and geographic effects; and (3) the effect of internal waves must be largely eliminated. If these conditions are attainable, the direction of relative currents can be established from bathythermograms. The results of this test indicate that the speed of the currents (0/305-dynamic meters) can be determined within 25 per cent of those obtained from Nansen bottle and reversing thermometer data (0/1000 dynamic meter).

Acknowledgments--The author desires to express his appreciation to members of the Crossroads Oceanographic Group in Support of Radiological Safety Reconnaissance, especially to R. Revelle, C. A. Barnes, John Lyman, and Dean Bumpus, for making possible the collection of data for this report.

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(Manuscript received August 2, 1949; presented at the Twenty-Eighth Annual Meeting, Washington, D. C., April 28, 1947; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

ZOOPLANKTON AS AN INDEX OF WATER EXCHANGE BETWEEN BIKINI LAGOON AND THE OPEN SEA

Martin W. Johnson

(Contribution from the Scripps Institution of Oceanography, New Series, No. 413)

Abstract--The zooplankton (small floating animals) are useful indicators of water currents in the area where they live. Of the holoplanktonic (permanently planktonic) forms found in Bikini Lagoon, 90 per cent or more occur also in the surrounding open sea. The lagoon plankton, however, is characterized by (1) a number of endemic species not found outside the lagoon except where washed out by outflowing currents, and (2) a concentration within the lagoon of many forms that occur normally also in the outside surface waters. Both of these features are dependent upon the degree of dispersal of the plankton and are therefore directly correlated with the semiclosed circulation which the lagoon imposes upon the water that flows in over the eastern and northeastern reefs.

A study of the vertical distribution and diurnal vertical migration of the endemic copepod *Undinula vulgaris*, and of the general plankton, reveals that much of the lagoon plankton lives mainly in the deeper slowly (0.1 knot) eastward-moving counter current which as a water mass represents 70 to 90 per cent of the living space in the lagoon. Even during the night when much of the plankton migrates towards the surface into the faster (0.3 knot) westward-moving water, about two-thirds of the plankton still remains within the deeper counter current.

Much of the oceanic plankton swept into the lagoon thrives there and becomes concentrated so that the average concentration per cubic meter of water of the eleven most common animal groups is about four times higher than in the outside north-eastern area, and twice higher than the outside western-southern area of principal outflow. The increased concentration found in the outside water to the west and south indicates an outwash from the lagoon in that area and is shown especially by such forms as *Undinula vulgaris* and *Tretomphalus planus* (foraminifer), which occur abundantly in the lagoon, but were not found in the area immediately to the east and north of the atoll.

The restricted water flow within the lagoon is biologically important in conserving also the transitory planktonic larval stages of lagoon and reef fishes and bottom-living animals.

The presence of *Copilia mirabilis* (copepod) mainly in the southern half of the lagoon and outside around the southern and south-eastern periphery of the atoll suggests a slow flow of southern water into the lagoon between Enyu and Bikini Islands and through the eastern part of Enyu Channel.

By the term plankton is meant all of the organisms, usually microscopic or semi-microscopic that float about in the sea with little or no resistance to water currents. Thus, in general, they are dispersed and transported in a manner dependent upon the prevailing current system. The animal plankton consists not only of such forms as are permanently planktonic all of their lives (holoplanktonic), but also of the transitory young or larval stages of bottom-living forms, such as crabs, certain shrimp, clams, and others, and of the fishes.

In most planktonic studies it is customary to apply the results of hydrographic surveys as an aid to explaining the biology of an area. The mechanism presented by the current system is a factor of prime importance in production, concentration, and dispersal of the planktonic life. This dependence upon currents, therefore, sometimes makes it possible to take the converse point of view in order to gain information and corroborative evidence on the nature of water currents through a study of the plankton. The countless myriads of floating forms are in a sense tiny drift bottles, which, if recognized as to kind and source, are useful indicators. Similarly the concentration of a population presupposes a certain type of circulation that minimizes dispersal.

Perhaps 90 per cent or more of the holoplanktonic forms found in Bikini Lagoon occur also in the outside surrounding water. The lagoon plankton fauna are, however, characterized by two strik-

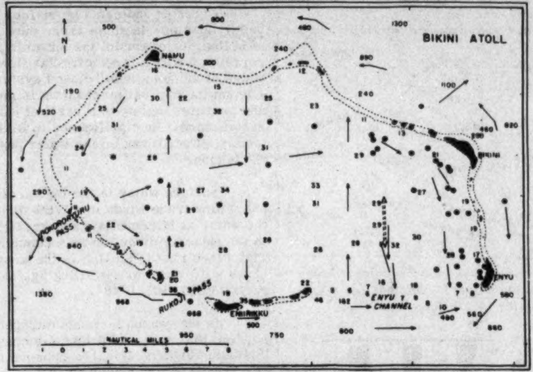

Fig. 1--Positions of plankton hauls taken at Bikini Atoll March 10 to May 18, 1946; the arrows indicate course of horizontal hauls; the circles indicate vertical hauls; the figures are depths in fathoms

ing features: (1) A number of the important species are endemic to the lagoon and are seldom or never found outside except at localities where they have been washed out of the lagoon over the reef or through passages. (2) There is a concentration of many forms that also occur normally in the outside surface water. In general the plankton concentration in the Marshall Island area is relatively very sparse.

The capacity of the lagoon to foster an endemic fauna, and the ability of the Lagoon to concentrate the fauna common to the Lagoon and the outside water, result largely from the same mechanism. This mechanism is inherent in the current pattern which results mainly from the retarding effect that the lagoon has on the oceanic water which it intercepts from the North Equatorial Current.

Bikini Atoll is a coral-capped volcanic mountain which rises abruptly from the ocean floor at a depth of over 2000 fathoms. The Lagoon has a maximum depth of about 35 fathoms [EMERY, TRACEY, and LADD, 1949]. At low tide it is completely isolated from the sea except for a number of passages forming gaps of six to 35 fathom in depth in the reef along the southern and southwestern periphery (see Fig. 1).

Methods-- The plankton samples forming the basis for this report were collected with Clarke-Bumpus plankton samplers, which give a measure of the volume of water filtered through the net. The nets used with these samplers for the regular series of hauls were of No. 2 bolting silk. In this series 64 samples are from inside the lagoon and 54 outside the lagoon.

Inside the lagoon, two samplers were usually towed 20 to 30 minutes sumultaneously at one- and ten-fathom depths. Deeper horizontal towing within the lagoon was inadvisable because of the many coral heads or pinnacles rising from the bottom. For the 24-hour series in the lagoon three samplers were towed at 1-, 10-, and 25-fathom depths over a course that had previously been carefully surveyed to determine that no coral obstructions would be encountered. Outside the lagoon, the samplers were towed simultaneously at 1-, 10-, and 40-fathom depths. The positions and courses over which the samples were taken are shown in Figure 1.

The regular hauls were checked qualitatively by many irregular qualitative samples collected with larger nets towed at various places inside and outside the lagoon at depths up to 260 fathoms.

This report deals mainly with observations made between March 10 and May 18, 1946, before the atomic bomb blasts. Samples taken after the blasts (not shown in Fig. 1) were concerned principally with conditions within the lagoon. The few samples taken outside as late as August 8, 1946, and contrasted with the spring samples, showed no appreciable change in concentration in the plankton.

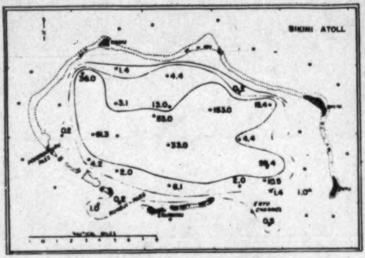

Fig. 2--Undinula vulgaris, number per cubic meter of water by stations (black dots)

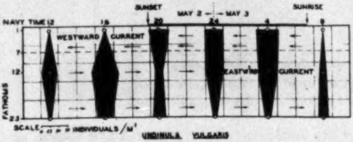

Fig. 3--Diurnal migration of Undinula vulgaris; thirty-five-minute horizontal hauls taken at surface, ten, and 25 fathoms at four-hour intervals over course shown by double dashed arrow in Figure 1

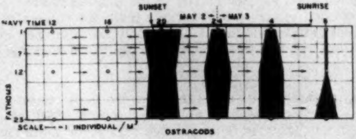

Fig. 4--Diurnal migration of Ostracods

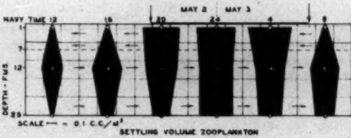

Fig. 5--Diurnal migration of total zooplankton as represented by settling volume

Currents indicated by maintenance of endemic species--In some areas such as the Gulf of Maine, for example, the endemic population is correlated with a large circular flow of water which produces a semi-closed system. In Bikini Lagoon the current mechanism is much smaller and operates mainly in a vertical rather than horizontal plane. Its significance in Bikini is the retarding effect it has on the water passing through the lagoon.

Surface water in the lagoon is driven by the prevailing trade winds in a WSW direction, whereas the water at intermediate and greater depths flows in the opposite direction as a counter current. This gives rise to sinking in the western portion of the lagoon and to upwelling against the eastern reefs [VON ARX, 1948].

In the search for endemic species in Bikini Lagoon, the most useful thus far considered are Undinula vulgaris (Dana), a copepod crustacean, and Tretomphalus planus Cushman, a pelagic foraminiferan. The latter was identified by J. A. Cushman who described the species originally from Samoa, and states that it occurs also at Fiji Islands. Both occur also in the lagoons of Eniwetok, Rongelap, and Rongerik.

Undinula vulgaris is a widely-distributed species and though reported from the open ocean it appears to be preponderantly a lagoon form in this area. Large (two millimeters) and easily recognized, it occurs in varying numbers throughout most of the Lagoon. During the period March 10 to May 18 it was present at about 90 per cent of the lagoon stations. The largest numbers (up to 255 per cubic meter of water) have been taken near the center of the lagoon (see Fig. 2). In the daytime the animals seek the deeper water strata, ten or more fathoms deep. Very few specimens were caught in the net towed at or above the one-fathom depth except at night. More will be said about this later. It is obvious that to maintain an endemic population of an amimal with a relatively long life history like that of Undinula vulgaris, and in such numerical concentrations, there must be a pattern of currents which leads to a relatively closed system with very slow flushing of at least a large part of the Lagoon. This would require the presence of a sizable counter current or some other effective retarding system. The hydrographic study has indicated that a deep slowly moving counter current exists. Thus the biological and hydrographic observations support each other on this point.

During the time that the animals are in the upper surface layer near the outwash of the Lagoon, they are in danger of being swept out through passages or over the reef. Therefore, (a) reproduction in the deeper layer would have to be at a rate sufficient to replenish the loss of numbers washed away by the surface layer, or (b) in case of slow reproduction the bulk of the population would have to spend the greater part of the time in the deep slowly-moving counter current. That the latter occurs in Bikini Lagoon is indicated by a study of the diurnal vertical migratory journeys of Undinula vulgaris and a number of other planktonic forms (see Fig. 3, 4, and 5). From these figures it will be seen that not only Undinula but the bulk of the other planktonic forms avoid the more

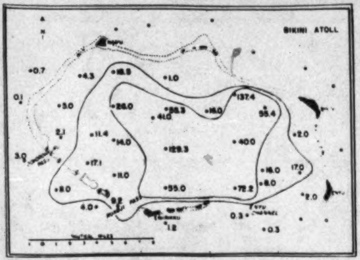
N
BIKINI ATOLL
0.7
4.3
18.5
1.0
0.1
3.0
26.0
36.3
41.0
16.0
137.4
95.4
2.1
11.4
14.0
129.3
40.0
2.0
3.0
17.1
11.0
16.0
17.0
8.0
8.0
55.0
72.2
2.0
9.2
4.0
0.3
1.2
0.3
NAUTICAL MILES

Fig. 6--Tretomphalus planus, number per cubic meter of surface water by stations (black dots)

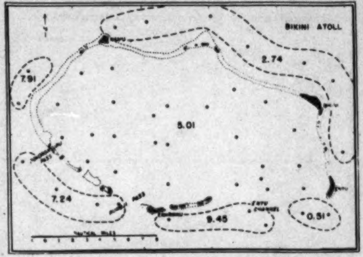
N
BIKINI ATOLL
2.74
7.91
5.01
7.24
9.45
0.51
NAUTICAL MILES

Fig. 7--Globigerina, average number per cubic meter of water by areas; black dots indicate stations averaged in each area

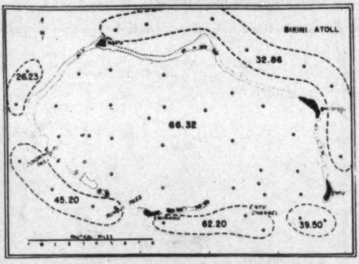
N
BIKINI ATOLL
32.86
26.23
66.32
45.20
62.20
39.50
NAUTICAL MILES

Fig. 8--Copepods, average number per cubic meter of water by areas; black dots indicate stations averaged in each area

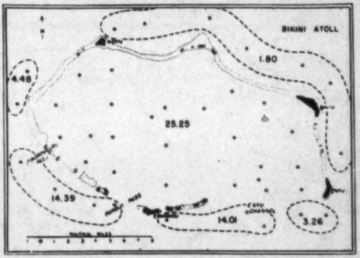
N
BIKINI ATOLL
1.80
4.48
25.25
14.39
14.01
3.26
NAUTICAL MILES

Fig. 9--Appendicularia, average number per cubic meter of water by areas; black dots indicate stations averaged in each area

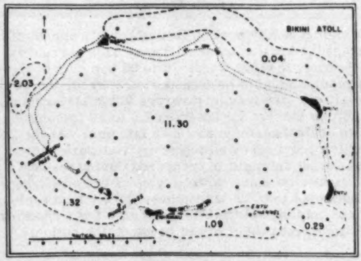
N
BIKINI ATOLL
0.04
2.03
11.30
1.32
1.09
0.29
NAUTICAL MILES

Fig. 10--Pteropod, Creseis sp., average number per cubic meter of water by areas; black dots indicate stations averaged in each area

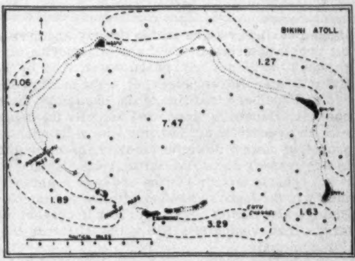
N
BIKINI ATOLL
1.27
1.06
7.47
1.89
3.29
1.63
NAUTICAL MILES

Fig. 11--Arrow worms, Sagitta spp., average number per cubic meter of water by areas; black dots indicate stations averaged in each area

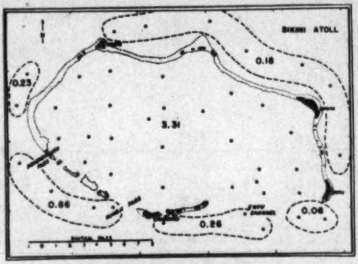

Fig. 12--Crab larvae, average number per cubic meter of water by areas; black dots indicate stations averaged in each area

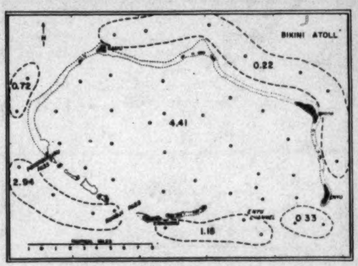

Fig. 13--Shrimp larvae, average number per cubic meter of water by areas; black dots indicate stations averaged in each area

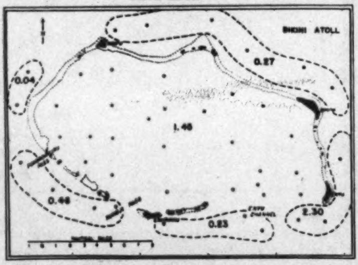

Fig. 14--Fish eggs, average number per cubic meter of water by areas; black dots indicate stations averaged in each area

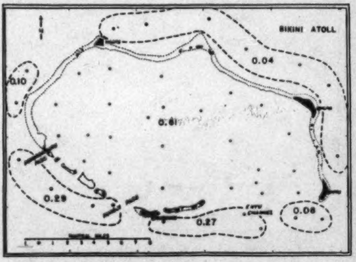

Fig. 15--Fish larvae, average number per cubic meter of water by areas; black dots indicate stations averaged in each area

lighted surface waters during the day and even at night a goodly portion (two-thirds or more) of the population remains in the slowly moving (average 0.1 knot) counter current. This current involves 70 to 90 per cent of the water mass of the Lagoon; in other words 70 to 90 per cent of the living space. Nevertheless, at night much of the population must be transported towards the western and southern margins of the Lagoon by the relatively fast-moving (average 0.3 knot) surface current. Depending upon wind velocity the depth of this current varies from 2.5 to 10 fathoms, with an average of 6.5 fathoms (the maximum depth of the Lagoon is about 33 fathoms). At the approach of dawn a downward movement of the animals occurs and a slow journey eastward begins. In this manner about one-third of the distance lost during the night is recovered during the day. It is probable that this is the order of events with respect to many larval forms of bottom-living adults in the lagoon and also with the oceanic forms swept into the lagoon over the eastern reefs. Biologically this is very important in conserving not only the holoplankton, but also the transitory larvae that eventually settle to the bottom of the lagoon or on the reefs to re-stock the bottom-living animals.

Tretomphalus planus was found in all parts of the lagoon, sometimes in numbers up to 175 or more individuals per cubic meter of water. The average was, however, only about 25 per cubic meter. It was present throughout the whole period of investigation (March 10 to Aug. 8) and occurred at all depths within the lagoon. The greatest concentrations were in the upper ten fathoms except in a few instances where numbers were greater in the deeper strata.

The plankton samples collected outside the lagoon yielded this species only at the stations situated along the western and southern periphery of the atoll. The greatest outside concentrations were found near the passages, thus suggesting the main route of outwash for the species as indicated in Figure 6. Apparently small numbers are also washed out over the long shallow reef extending from Bokorōryūru Pass to Namu Island. This is in keeping with the animal's habit of floating near the immediate surface. The deeper-living lagoon animals such as *Undinula vulgaris* are discharged through the passages (depths varying from six to 35 fathoms) rather than over the reef.

From the above observations it appears that *Tretomphalus* is endemic to the Lagoon, perhaps requiring shallow water during the early stages of its life history before the float chamber has been developed. This would explain its absence or scarcity in deep water outside the lagoon except where washed out on the down-current side by the prevailing flow through the lagoon. It should be pointed out here that this relationship suggests the use of *Tretomphalus* as an index of past oceanographic conditions of the atoll. Provided its tests are well preserved as fossils in the sediments, it should serve to show the general direction of currents that have swept through the Lagoon in past history, for though the animal is kept suspended by the float chamber while alive, there must be a gradual precipitation of defunct tests to the bottom along the course of the outflowing water.

Maintenance and concentration of oceanic plankton in the Lagoon--The salinity and temperature of the water within the lagoon differs but little from that outside. The highest lagoon salinity (34.85 °/oo) is only about 0.1 °/oo higher than in the open ocean while the temperature change from about 27.02 to 27.10° C is negligible. The vertical gradients of these factors are also insignificant. In these respects the lagoon water is essentially oceanic and would therefore not be seriously inimical to most of the surface oceanic plankton coming in over the reef.

The concentration of oceanic plankton that has drifted into the Lagoon is a further indication of a semi-closed circulatory pattern in the Lagoon. Presumably this circulation would operate in a mechanical manner similar to its effect on the maintenance of the endemic fauna. In the following section it will be shown that certain forms that are common to the inside and outside may be as much as 27 times more numerous per unit of water inside the lagoon. This increased concentration doubtless results in part from better food conditions inside as a result of local production, but the reduction of dispersal must be very important.

The presence in the Lagoon of many egg-bearing specimens and a large proportion of larval and juvenile stages of species common to the outside is evidence of successful reproduction inside. In the absence of wide and uninterrupted dispersal, such as occurs outside, a concentration from both reproduction inside and recruitment from outside should occur.

In summarizing the average concentration by areas of the eleven most common animal groups, it appears that the ratios of concentration per cubic meter for (a) the outside eastern-northern area, (b) the inside lagoon area, and (c) the outside western-southern area of principal outflow, are roughly a:b:c: = 1:4:2.

Inflow and discharge of water from the Lagoon as indicated by the drift of planktonic life--When we compare the number of specific organisms occurring per cubic meter of water at stations occupied inside the lagoon with the number found in the surrounding water just outside the reefs and passages, it is seen that the concentration is almost invariably smallest in the waters outside the eastern and northern reefs. It is nearly always largest by far at the stations within the Lagoon, and intermediate at stations just outside the western and southern passages and reefs. The largest outside catches occur near Rukoji Pass and the western portion of Enyu Channel (Figs. 2, and 6 to 15). Figures 2 and 6 are especially useful since they deal with two endemic species and show clearly the route by which the species are washed out of the Lagoon. According to hydrographic studies the outflow is through these passages and over these reefs, and this is clearly corroborated by the biological evidence. The biological evidence does, however, indicate a somewhat broader outwash from Enyu Channel in the early part of the survey than the hydrographic charts indicate.

From the hydrographic survey we learn that in the open sea off Bikini there is a small north-south gradient in temperature and salinity so that the water entering the lagoon from the north is about 0.1 °/oo C colder and 0.3 °/oo more saline than that which enters the southern passes.

The plankton has not yet been studied in sufficient detail to know the extent of its usefulness in indicating other than the general routes of entry into the lagoon. The distribution of one copepod species, *Copilia mirabilis*, does, however, suggest that its center of distribution may be in the southern water with which it enters the lagoon through Enyu Channel and over the reef between

Bikini and Enyu Island. Forty-four per cent of the stations in the southern half of the lagoon yielded this species, but only 11 per cent in the northern half. Outside the lagoon from near Bikini Island around the southern periphery of the atoll to near Arriikan Pass on the west side, 75 per cent of the stations yielded the species. None was taken at outside stations north of these points. Nowhere was the species found in abundance.

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VON ARX, W. S., The circulation systems of Bikini and Rongelap Lagoons, Trans. Amer. Geophys. Union, v. 29, pp. 861-870, 1948.

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La Jolla, California

(Manuscript received August 2, 1948; presented at the Twenty-Eighth Annual Meeting, Washington, D. C., April 28, 1947; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

ORGANIC PRODUCTIVITY OF AN ATOLL

Marston C. Sargent and Thomas S. Austin

(Contribution from the Scripps Institution of Oceanography, New Series, No. 411)

Abstract--The production of organic matter by phytoplankton in the waters in and around the northern Marshall Islands is extremely low. The amount of plankton swept across the reefs of these islands by the wind-driven currents is grossly inadequate to support the animals living on the reefs. The reef community, consisting of attached algae and animals, is self-supporting; all the algae produce at least as much organic matter as all the animals consume. The rate of production of organic matter by an atoll as a whole - water, reef, and bottom - is several times as high as that of the surrounding sea, and permits maintenance of a marine population considerably denser than that of the ocean. The maximum possible rate of growth of the reef is found to be 1.4 cm per year.

Measurements of the oxygen and phosphate content of the waters in and near some of the atolls of the northern Marshall Islands yield evidence on how these coral structures remote from any terrigenous influence maintain themselves both as geographical features and as biotic communities supporting marine populations considerably denser than those of the surrounding waters. Conclusions of this kind can be drawn from such measurements because of the fact that, in the formation of organic matter, oxygen is produced and phosphate consumed, while in the biological combustion of organic matter these processes are reversed. Thus, during a time interval, the appearance of oxygen or the disappearance of phosphate in an identifiable body of water is a measure of the production of organic matter.

The recent comprehensive review by YONGE [1940] of the biology of corals makes any partial review of earlier work by the present writers unnecessary and indeed supererogatory. SHEPARD [1948, pp. 251-278] has reviewed the geology of atolls.

The standing population of plankton inside and especially outside the lagoons of the northern Marshalls is very small. The average depth at which the Secchi disc disappeared in the open sea was 41 m. According to the data of UTTERBACK [1936] see also SVERDRUP, JOHNSON and FLEMING [1942, pp. 82-84], this is representative of the clearest oceanic water and indicates a very small content of suspended matter. The average Secchi disc reading in Rongelap Lagoon was 18 m. While oceanic water is generally more transparent than this, waters as shallow as the lagoons, with an average depth of 49 m, are seldom so transparent. The conclusion that the standing population is small is supported by data on biochemical oxygen demand obtained by Donald B. Johnstone and on plankton catches obtained by Martin W. Johnson.

The rate of production of organic matter in the surface water was measured by the difference in oxygen content between bottles kept in the light and dark for eight or 24 hr. At eight stations outside Rongelap Atoll (see Table 1) the mean difference was less than 0.01 ml per liter which is equivalent to less than 0.005 mg of carbon per liter fixed as organic matter per day. The mean production of oxygen in 24 hr was 0.04 ml at eleven stations in Eniwetok Lagoon and 0.01 ml at seven stations in Rongelap Lagoon, equivalent to 0.02 and 0.005 mg of carbon respectively. In view of the large standard deviations at most stations the apparent difference in the rate of production between Eniwetok or Rongelap Lagoon and the open sea cannot be considered significant. Compared with the measured rate of production in sea water in other parts of the world all of these values are small [see RILEY 1941, p. 84, reproduced in SVERDRUP, JOHNSON and FLEMING, 1942, p. 938].

On the eastern reefs of Rongelap Atoll the ocean water, driven steadily across the reef from sea to lagoon by the trade wind waves, undergoes readily measurable changes in its oxygen and phosphate content because of the activities of the sedentary reef organisms. The presumptively productive section of a reef extends from the depth of the compensation point on the seaward slope of the reef (the sedentary animal community goes even deeper) upward through the surf zone and across the reef top to whatever point is selected as marking the inner edge of the reef, and beyond which the bottom is considered part of the lagoon. Only between the surf zone and the inner edge of the reef was the current steady enough in direction to make measurements possible.

Table 1--Oxygen production in surface ocean waters near Rongelap Atoll

Date	Location	Light condition	Oxygen content after eight-hour exposure				Average	Daily oxygen production
1946			ml/l	ml/l	ml/l	ml/l	ml/l	ml/l
June 24	Five miles NE of Kabelle Island	light	4.23	4.21	4.19	4.18	4.203	0.02
		dark	4.19	4.18	...	...	4.185	
June 24	Ten miles NE of Kabelle Island	light	4.21	4.20	4.20	...	4.203	-0.01
		dark	4.21	4.21	...	...	4.210	
June 24	Fifteen miles NE of Kabelle Island	light	4.22	4.20	4.20	...	4.207	0.00
		dark	4.22	4.20	...	...	4.210	
June 25	Two miles S of South Pass	light	4.36	4.35	4.33	...	4.347	0.03
		dark	4.30	4.30	...	...	4.315	
June 25	Two miles S of Pigan Pass	light	4.31	4.30	4.28	...	4.297	-0.06
		dark	4.37	4.35	...	...	4.360	
June 25	Two miles S of Kaeroga Pass	light	4.34	4.33	4.30	...	4.323	0.01
		dark	4.33	4.30	...	...	4.315	
July 29	Five miles E of Kabelle Island	light	4.32	...	...	...	...	0.01
		dark	4.31	...	...	...	...	
July 30	Six miles W of West Pass	light	4.36	...	...	...	...	0.07
		dark	4.29	...	...	...	...	
							Mean	0.01

For the purpose of making these measurements, four stations were selected, spaced equidistant from the surf zone across to the inner edge of the reef north of Bokujarito Island, Rongelap Atoll. The depth, velocity, and temperature were measured and oxygen and phosphate samples taken at each station approximately every three hours, tide permitting. During the period of high tide, it was impossible to work on the reef. From the data thus obtained (Table 2) the product of (a) the velocity, (b) the cross section, and (c) the difference in oxygen content between the surf zone and the inner edge of the reef can be computed and gives the total oxygen consumed or produced by the reef section between the limits of the sampling.

A second series of measurements at the same site and a series at another location on the eastern reef of Rongelap Atoll gave accordant figures.

The reef community during the night low tide consumed 0.22 ml of oxygen per sec per cm of reef measured perpendicular to the direction of the current (see Table 3), which amounts, on the average, that is, at half tide, to consuming 0.16 ml of oxygen from each liter of water flowing over the reef. This amounts to one-sixth of the oxygen removed by micro-organisms from each liter of oceanic sea water in several days or a week during the biochemical oxygen demand measurements (see Fig. 1). As the water spends only ten to 20 min in crossing the reef, it is unlikely that filtration is efficient enough to remove even ten per cent of the nutrient matter which bacteria can oxidize in a week. The plankton caught with a No. 8 net corresponds to only 0.3 per cent of the organic matter consumed by the reef. The reef organisms, therefore, can hardly be supported by the organic matter brought to them from the open sea and must be dependent on a portion of that synthesized on the reef.

From observations made at different times of day, and by using some reasonable assumptions about the rate of photosynthesis at different times of day, it is possible to construct a curve showing the rate of oxygen exchange and hence of organic matter metabolism by the reef throughout 24 hr. Integration of the curve thus obtained (see Fig. 2) indicates that during a whole day there was a small excess of oxygen production over oxygen consumption. As this excess is only about 25 per cent of the total oxygen produced during the daylight hours, its absolute value is subject to considerable uncertainty. However, it is clear that on the reef top, between the surf zone and the inner edge, the daily production of organic matter by the photosynthetic organisms at least equalled the consumption of organic matter by all the organisms. The gross daily production of organic matter by the reef top (equal to the daily apparent photosynthesis plus the concurrent respiration) was 22 liters of oxygen or 12 gm of carbon per cm perpendicular to the direction of flow of current. On the 300-m wide reef this amounts to 4.0 gm/m^2 of carbon per day.

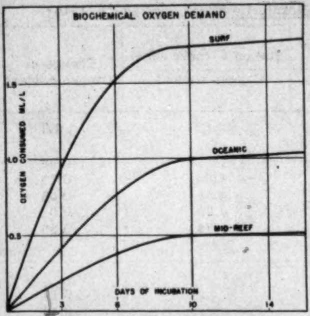

Fig. 1--Biochemical oxygen demand; data furnished by Donald B. Johnstone

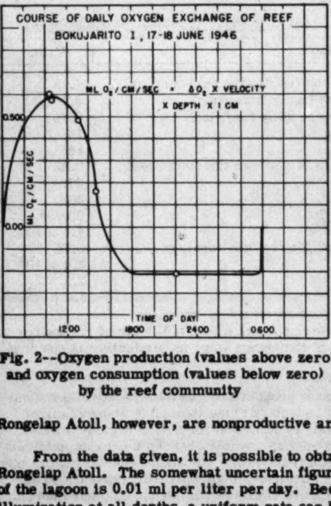

Fig. 2--Oxygen production (values above zero) and oxygen consumption (values below zero) by the reef community

The density of the population of algae, mostly nullipores, on the seaward face of the reef in and outside of the surf zone, leads to the presumption that the productivity of that area is high. The biochemical oxygen demand of ocean water and of water from the middle of the reef were both appreciably less than that of the surf zone water. As the surf breaks over the face of the reef, organic matter is washed from the reef and carried a short distance out to sea, to be again hurled against the reef with the next breaking wave. Tentatively, the conclusion may be drawn that the whole reef community from the seaward slope inward produces even more organic matter than shown by the direct measurements, which were made only on the inner portion of this community.

The data on biochemical oxygen demand (Fig. 2) are not consistent with the data on oxygen consumption by the reef (Table 3). According to the former, organic matter equivalent to more than 1.0 ml of oxygen per liter disappeared from the water between the surf zone and mid-reef. According to the latter, only 0.16 ml of oxygen was consumed from each liter of water as it crossed the reef. We are unable to explain this discrepancy to our entire satisfaction.

The higher portions of the reef, represented by the tide pools, may be considered as relatively sterile. The maximum rate of oxygen consumption in such a pool was only about 1/50 that of the densely populated areas between the islands where the circulation of water is good. The bottoms of these pools contain a few corals, many of which are dead and covered with a great number of foraminiferal tests. The rate of oxygen production during daylight hours, in the pools studied, was not sufficient to balance the consumption of oxygen during 24 hr. Orr and Moorehouse [ORR, 1933], during the course of the Great Barrier Reef Expedition, made similar diurnal studies on tide pools isolated from the oceanic circulation. In these pools the living coral were very abundant, and the changes which they observed they attributed almost entirely to these corals and their associated zooxanthellae. Their data indicate that during a 24-hr period there was a net oxygen production. The pools on the reef at Rongelap Atoll, however, are nonproductive areas.

From the data given, it is possible to obtain a crude estimate of the total organic production of Rongelap Atoll. The somewhat uncertain figure for gross oxygen production in the surface waters of the lagoon is 0.01 ml per liter per day. Because of the rapid vertical mixing and the considerable illumination at all depths, a uniform rate can be assumed throughout the $4 \times 10^{10} m^3$ of water in the lagoon. The total daily production is, therefore, of the order of 0.2×10^9 gm of carbon per day fixed as organic matter. This rate, restated as 0.2 gm/m^2 of carbon per day (over the $810 \times 10^6 m^2$ of lagoon), can be taken as close to the rate of organic production in the ocean waters surrounding the atolls. The total daily production of the atolls is larger first because of the reef community of photosynthetic organisms and second because of the bottom community. The total production of the $28 \times 10^6 m^2$ of reef can be computed as 0.1×10^9 gm of carbon per day. Data are not available for the productivity of the lagoon bottom, but the population of photosynthetic organisms is relatively sparse, and it is likely that the contribution of each square meter of the bottom is much less than that of the reef. The oxygen content of the bottom waters of the lagoon, because of the large volumes of water involved, would probably not be a sensitive indicator of bottom activity. In any case there is no measurable daily cycle in the oxygen content of the bottom waters. The total production of the atoll is,

Table 2--Stations on section across reef -- Bokujarito Island, June 17-18, 1946

Date and time	Station 1 (surf zone)			Station 2	Station 3	Station 4 (inner edge)	Change in Oxygen content
	Oxygen content	Depth	Velocity	Oxygen content	Oxygen content	Oxygen content	
June 17	ml/l	cm	cm/sec	ml/l	ml/l	ml/l	ml/l
10^h20^m	4.21	18	35	4.85	5.19	. . .	0.98
11^h50^m	4.25	. .	. .	4.50	4.77	4.80	0.55
14^h40^m	4.28	64	84	4.34	4.31	4.31	0.03
17^h50^m	4.31	. .	. .	4.23	4.27	4.29	-0.02
22^h00^m	4.09	18	35[b]	4.03	3.94	3.74	-0.35
June 18							
01^h00^m	4.19	. .	. .	4.27	4.02	. . .	-0.17
04^h00^m	4.26[a]	. .	. .	. . .	4.25[a]	. . .	-0.10
10^h30^m	4.49	20	73	4.57	4.67	4.88	0.39

[a] Position approximate [b] Estimate

Table 3--Oxygen consumption and growth rate of reef at Bokujarito Island, Rongelap Atoll

Item	Description	Value
1	Oxygen content of water flowing onto reef top just inside surf zone, ml/l	4.09
2	Oxygen content of water leaving reef top on lagoon side, ml/l	3.74
3	Decrease in oxygen content of water while crossing reef, ml/l	0.35
4	Depth of water, cm	18
5	Velocity of current, cm/sec	35
6	Oxygen consumption of reef top, ml/sec/cm of reef normal to current	0.22
7	Width of reef, m	300
8	Oxygen consumption of reef top, $ml/m^2/hr$	262
9	Average rate of coral respiration, ml/gm/hr	0.0059
10	Average coral population of reef top, gm/m^2	4.4×10^4
11	Average growth rate of coral, gm new coral formed/gm coral/yr	0.76
12	Growth rate of reef, $gm/m^2/yr$	3.4×10^4
13	Average density of coral, gm/cm^3	2.5
14	Growth rate of reef, cm/yr	1.4

then, greater than the sum of the lagoon water and reef production (0.3×10^9 gm of carbon per day) but less than what would appear if the whole bottom of the lagoon were as productive as the reef ($810 \times 10^6/m^2/day = 3 \times 10^9$ gm C/day).

It can be suggested that the primary reason for the productivity of atolls relative to the surrounding waters is that on these elevated structures sedentary algae (including zooxanthellae) find a footing in the upper water zones where light intensity is sufficient for active photosynthesis. Rather than filtering organic matter, the atoll community as a whole absorbs inorganic nutrients from the passing Equatorial Current. This suggestion rests on the view that sedentary algae utilized the available supply of inorganic nutrients more effectively than a plankton population can, but is not meant to imply that the algae of atolls are more effective than those of other shores.

From measurements of the oxygen consumption by the whole reef top in the dark can be derived an estimate of the maximum possible rate of growth of the reef. This can be done by dividing the measured oxygen consumption of the reef by the average of known rates of coral respiration at the same temperature and thus arriving indirectly at an estimate of the amount of living coral on the reef. From this estimate, by means of published growth rates of coral, it is possible to compute the growth rate of the reef. The advantages of the method are first that it avoids the almost insuperable difficulties of sampling and measuring to arrive at a direct measurement of the reef population and second that it estimates coral growth from a process with which it is undoubtedly closely linked, that of coral respiration. The errors of measurement of coral respiration pointed out by YONGE [1937] do not enter here, because both our measurements and the published ones were made in the same way. The greatest error in our method lies in the assumption that the total respiration of the reef is identical with the respiration of the lime-depositing organisms. Although other types of organisms do not constitute a large part of the mass of the community, some, like bacteria, have respiratory rates many times as large as that of the corals and may account for the greater part of the oxygen consumption by the reef.

The data used are shown in Table 3. Items 1 to 8 are direct consequences of the fact that water flows steadily in one direction across the reef. The average rate of coral respiration, item 9, is based on data from YONGE, YONGE and NICHOLLS [1932, p. 218]. Data published by MAYOR [1924a, p. 19] and VERWEY [1931, pp. 175-176] as well as our own direct measurements of coral, respiration are accordant with these. The average growth rate of coral item 11, is calculated from data of MAYOR [1924b, p. 62-64]. The average density of coral (item 13) was determined by K. O. Emery.

The growth rate thus computed is believed to be larger than the actual rate, but it is of the same order of magnitude as estimates obtained by other methods and to the extent that it is independent enhances their credibility.

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(Manuscript received October 20, 1948; presented at the Twenty-Eighth Annual Meeting, Washington, D. C., April 28, 1947; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

NOTE ON VOLCANIC ROCKS, WITH SPECIAL REFERENCE TO PLAGIOCLASE FELDSPARS, FROM MT. BOGANA, BOUGAINVILLE ISLAND, SOLOMON ISLANDS

George Baker

Introduction

Little is known of the petrology of the Quaternary-Recent volcanoes on Bougainville Island. The purpose of this article is to record the nature of ejectamenta and lavas in the Mt. Bogana-Torokina River area. Suggestions have been made herein concerning the origin of a mixed assemblage of non-zoned, variously zoned, and twinned andesine phenocrysts contained in the lavas.

The described rocks consist of ejected pebbles and andesitic scoria that form the greater part of the recent, symmetrical volcanic cone of Mt. Bogana (see Fig. 1), and slightly older andesitic flows that occur to the southwest in the valley of the Torokina River.

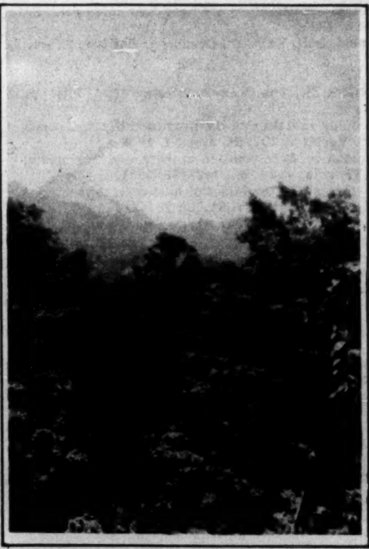

Fig. 1--Quiescent phase of Mt. Bogana, viewed from the southwest [Official Army War Photograph No. MH91169]

Mt. Bogana is situated near the center of Bougainville Island (see Fig. 2), the largest of the British Solomon Islands. Bougainville is 415 mi northeast of Papua, and 200 mi east of New Britain, where the nearest Recent volcanoes, other than those in the Solomon Islands, occur.

Centers of volcanic eruption on Bougainville Island lie along an arc extending southeasterly from New Ireland, through the Solomon Islands to the New Hebrides. This volcanic arc continues southeasterly to New Zealand, and is referred to by STANLEY [1921, p. 8] as lying along a structural unit called the Second Australian Arc, a double arc uniting in Bougainville Island.

The two known, recently active volcanoes in Bougainville Island, are Mt. Bogana in the Crown Prince Ranges, and Mt. Balbi [FISHER, 1939, pp. 16 and 21] further northwest (see Fig. 2) in the Emperor Ranges. Both volcanoes have broken through gneisses which are said [STANLEY, 1922, p. 32] to form the basement rocks of Bougainville Island.

FISHER [1939, p. 16] reported that Mt. Bogana was more or less continuously active. The eruption of May 15, 1938 gave off considerable volumes of dust, and clouds of steam. Dust, pumice, and solid rocks accumulated over a large area southwest of the cone. These products partially covered up the earlier-formed andesites out-cropping along the Torokina River. When observed from June 11 to 15, 1945, the volcano was quiescent (see Fig. 1). Steam issued from the vent, and occasional small explosions ejected rock fragments from the crater. The ejection of material was accompanied by frequent slight earth tremors with occasional stronger tremors. The natives predicted that Bogana would erupt violently in 1946, but no great activity was reported during that year.

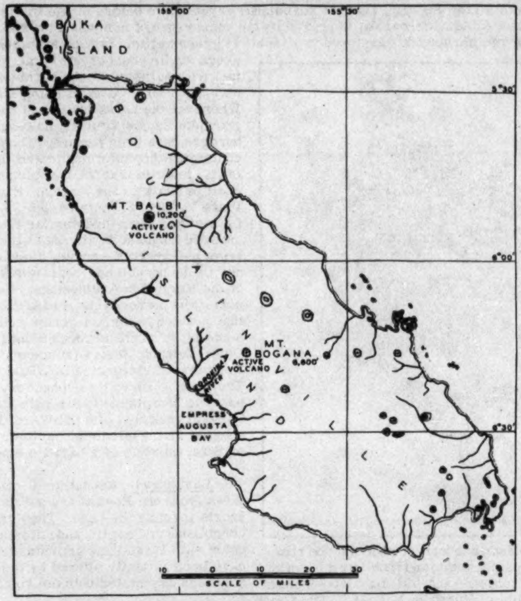

Fig. 2--Sketch map of Bougainville Island showing position of Mt. Bogana and the Torokina River [based on U. S. Army Maps SB56-3 and SB56-6; 1:500,000, first edition, 1942]

Recent volcanic products

Recent ejectamenta and occasional flows from the cone of Mt. Bogana. Lava flows from earlier eruptions occur along the Torokina River. The principal rock types among the lavas are porphyritic hornblende and augite andesites. Some of these flows have been altered metasomatically.

Ejected pebbles--Rounded rock fragments from Mt. Bogana represent rock torn from the walls of the volcanic conduit. Several rock types occur: (1) Serpentine rock with calcite veins, radial aggregates of quartz and pseudomorphs after olivine (the parent rock was apparently an olivine dolerite); (2) granophyric rock of fine-grained, interlocking quartz, orthoclase, and albite; (3) diabase partially replaced by calcium carbonate and containing residual augite and introduced pyrite; and (4) cryptocrystalline epidote rock.

Scoria--The scoria from the recent cone of Mt. Bogana consists of vesicular augite-hornblende andesitic basalt. It is dark grey in color. Large phenocrysts of green augite and brown hornblende, and small phenocrysts of hypersthene and zoned-twinned andesine and labradorite occur in a brown, glassy base containing abundant iron ore minerals. Numerous needle-like prisms of augite are widespread in the brown glass, and also occur in some of the andesine phenocrysts.

The augite phenocrysts have sharply marked crystal boundaries, and include large patches of iron ore minerals. The hornblende phenocrysts, however, have been partially resorbed. Most of them are surrounded with reaction zones containing small prismatic augite crystals and granular iron ore minerals (see Fig. 3). That the hornblende crystallized before augite in the Bogana augite-hornblende andesitic-basalt, is proved by the occurrence of compound crystals composed of a core of brown hornblende enveloped in a jacket of green augite. In such examples, and where green augite phenocrysts come into direct contact with hornblende, the hornblende has no reaction borders, contacts being sharply marked. Wherever the hornblende is in contact with groundmass constituents, however, reaction borders have been formed. These borders, containing diopsidic augite and iron ore minerals, indicate that the hornblende was not in equilibrium with the magma. EDWARDS [1938, Plate V, Figs. 2, 3, and 4, pp. 75, 78, and 79] figured and described similar idiomorphic basaltic hornblende altered at the rim to aegerine, aegerine-augite, diopside and iron ore in the hornblende-bearing alkaline lavas of the Kerguelen Archipelago. He also found numerous phenocrysts of basaltic hornblende that reacted to form aegerine and iron ore, in some of the alkaline rocks belonging to the Older Volcanic Series (Miocene or earlier Tertiary) in Victoria [EDWARDS, 1939, p. 89]. The restriction of these phenocrysts of altered basaltic hornblende to strongly differentiated types of rock, led him [EDWARDS, 1938] to suggest that it forms during the cupola stage of differentiation of a basaltic magma.

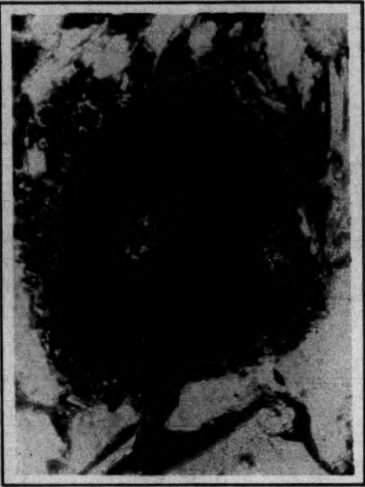

Fig. 3--Hornblende phenocryst with reaction rim containing diopsidic augite and iron ore minerals from augite hornblende andesitic basalt (×80, ordinary light) Mt. Bogana (photo by Miss M. L. Johnson)

Lava flows--Recent (or Pleistocene) lava flows from Mt. Bogana are not as basic as the scoria forming the cone. They are porphyritic hornblende and augite andesites of lighter color with crystalline groundmasses. They have been partially altered by the attack of solutions associated with the volcanic activity.

(1) Altered andesites--The metasomatically altered hornblende and augite andesites show various stages of chloritization, epidotization, albitization, and replacement by calcium carbonate, like some of the Upper Devonian meta-andesites described by HILLS [1932]. They are dense in texture and greenish-grey in color. Granular, secondary sphene and rare interstitial quartz occur in addition to the secondary minerals developed by the processes of metasomatism. Grains of iron-ore minerals are abundant. Pyrite is occasionally present along joint planes in the altered andesites.

Plagioclase phenocrysts, abundant in some of the altered andesites, are fresh and unaltered in some parts, and are elsewhere completely altered and cloudy. Under these circumstances, the andesine phenocrysts seldom show zoning and are partially replaced by calcite. Siderite also accompanies calcite in chlorite pseudomorphs after hornblende. Crystal outlines and occasional cleavage traces are all that remain to indicate that hornblende was the original mineral in these pseudomorphs. The groundmass of the altered andesites is chloritized and replaced by calcium carbonate in patches. Large apatite crystals occur occasionally. Epidote, zoisite, and clinozoisite occur in veinlets and rosette structures in the groundmass and as granules in some of the phenocrysts.

Sometimes, the plagioclase phenocrysts show selective replacement of certain of the polysynthetic twin bands. Some twin bands are clear and unaltered, but the alternate twin bands are replaced principally by chloritic matter. EMMONS and GATES [1943, p. 296], who have noted similar occurrences, consider that adjacent twin lamellae may not therefore be identical in anorthite content. They regard such phenomena [p. 288] as an indication that the twinning must antedate the incoming of the deuteric solutions causing the metasomatic alteration.

In parts of metasomatized augite andesite, unaltered grains of mauve-colored augite occur in a much altered groundmass with turbid andesine, patches of chlorite, occasional calcite and pyrite. The augite is apparently more resistant to chemical attack by deuteric solutions than the feldspars and hornblende.

The alteration of the Bogana andesitic lavas indicates abundant release of hydrothermal solutions during the end stages of volcanicity.

(2) Unaltered andesites--The unaltered augite andesites are light grey in color, with pale greenish-yellow augite phenocrysts and clear, colorless andesine idiomorphs. These are set in a fine-grained, felted groundmass of andesine laths, small augite prisms, and granules and clots of iron-ore minerals. Hypersthene is occasionally associated with the augite phenocrysts. The groundmass constituents sometimes show flow structure around the phenocrysts.

The unaltered hornblende andesites are light to dark grey. They contain green to brownish-green hornblende crystals, some of which show well-developed simple twins. Others show imperfectly developed zonal structures. With andesine, the hornblende comprises the greater number of the larger phenocrysts. Microphenocrysts of plagioclase feldspar are also present. The phenocrysts, occasional large apatite crystals and granules of iron-ore minerals, occur in a micro- to cryptocrystalline groundmass.

Marked and frequent changes in the composition of the hornblende andesite lavas are shown partly by the poorly zoned hornblende crystals, but more particularly by the andesine phenocrysts. The characteristics of the andesine phenocrysts are discussed in detail in the next section.

Andesine crystals in hornblende andesites

The characteristics of the andesine phenocrysts in the unaltered hornblende andesites are described in some detail, because in them, twinning, zoning, welding of crystals, resorbtion of crystals, and inclusions of earlier products of crystallization are best represented. Similar phenomena are by no means as well-marked in the augite andesites, the recent scoria or the altered andesites as the structures have been obliterated by metasomatic alteration.

Inclusions--Most of the unaltered andesine crystals are clear, but some contain inclusions of apatite, hornblende, iron-ore minerals, and indefinite, irregularly shaped particles. Some of these particles are isotropic, others are weakly birefringent and in optical continuity. The inclusions in the core of the crystal shown in Figure 4 are largely of this character, and are accompanied by inclusions of hornblende, apatite and iron-ore grains. Figures 5 to 7 show sketches of phenocrysts. Inclusions occur in various zones of the phenocrysts, but are more often confined to central cores (see Fig. 4 and 7). In twinned crystals (see Fig. 7L), the arrangement of inclusions in the cores is not affected by the twin boundaries. None of the outermost jackets of the zoned andesine phenocrysts contain inclusions.

Some phenocrysts contain feldspar inclusions (see Fig. 5A) entrapped in certain growth shells beneath the clear, outer jacket. They resemble examples described by HILLS [1940, p. 210], by HOMMA [1936b], and by WENK [1933]. These inclusions are in optical continuity and represent residual areas of almost completely resorbed, earlier-formed plagioclase. Like examples figured by HOMMA [1936a, Fig. 24, p. 154], the zonal inclusions are the result of incomplete reaction between the magma and the outermost zone of plagioclase at that particular instant of reaction. The reaction was initiated by changes in the temperature and composition of the magma. The partial resorption of this outermost zone undoubtedly had an important effect upon the composition of the next plagioclase zone which crystallized concomitantly with the reaction. Further changes in temperature and composition prevented complete diffusion, equilibrium was not reached, and the unresorbed material remained as inclusions. The process may have gone to completion in some zones at least which are free of inclusions.

The effect of incomplete resorption of earlier-formed plagioclase on the composition of zones is indicated in Figure 6G. The inclusions here are larger than usual, and lie across the zones.

Resorption--A number of andesine phenocrysts including zoned or twinned, or non-zoned or untwinned crystals show embayments which are filled by groundmass constituents. In some the embayments penetrate the core (see Fig. 7K) in others only the outer zones are embayed (see Fig. 6E). Apparently these embayments represent late-stage resorption when at times, and for various reasons, the phenocrysts were not in equilibrium with the magma.

Earlier stages of resorption during zonal growth are shown by the irregular contours of some inner chemical zones. A zone with a wavy outline is faintly discernible in the broad, light-colored outer jacket of the phenocryst shown in Figure 4. The same crystal possesses discontinuities at the angles and along some of the edges of inner zones. The inner cores of some phenocrysts have lost their euhedral outline (see Fig. 7I) because of extensive earlier corrosion. The twin structure and fabric of some partially resorbed andesine phenocrysts have been rebuilt and preserved by the subsequent growth of new plagioclase, similar to occurrences described in a Triassic basalt flow in Nova Scotia by FRIES [1939, p. 787].

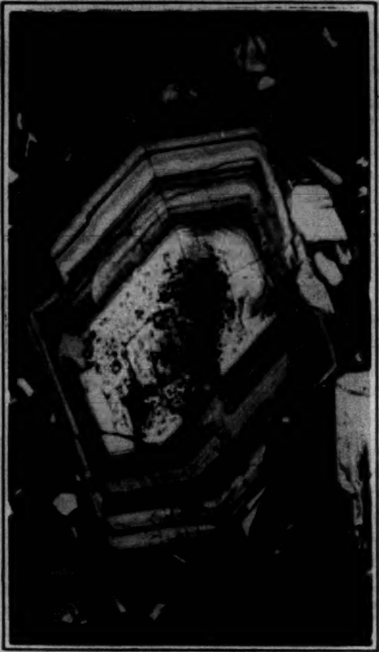

Fig. 4--Corroded, chemically zoned andesine phenocryst in fine-grained groundmass of hornblende andesite (× 44, crossed nicols), showing calcic core with inclusions enveloped by two main jackets, each with oscillatory zoning, from Torokina River, Bougainville Island (photo by Miss M. L. Johnson)

Welding of crystals--Small plagioclase crystals attached to the borders of phenocrysts of similar composition are common in the Bogana lavas (see Fig. 5A, 5D, 6H, and 7N). The welding-on of one plagioclase crystal to another has been described also by HILLS [1940, Fig. C, p. 212] and by EMMONS and GATES [1943].

The welding-on of plagioclase crystals (parallel and sub-parallel) to the margins of phenocrysts, occurred in the Bogana andesites during the formation of the outer zones of the phenocrysts. No crystals were observed attached to the core or to early-formed zones of plagioclase. If any were so attached, they have been completely resorbed. The welding-on process involves a linking up of neighboring crystals by the placement of the space lattices in parallel trends with little hindrance to the production of regular crystal forms during subsequent growth. The growth of later formed chemical zones in the larger phenocrysts led to partial enclosure of the smaller, marginal laths (see Fig. 5A, 5D, 6H, and 7N). Some of the marginal laths were zoned independently of and prior to the formation of the outer growth zones in the phenocrysts. During subsequent growth of the larger phenocrysts, there was little addition of new material to the smaller crystals, and little resorption along welded contacts between marginal laths and phenocrysts.

Several of the larger, more complex phenocrysts in the Bogana andesites, consist of three or four crystals that have been welded together. This phenomenon is also noted by FRIES [1939, p. 785]. In the various parts of these crystals, chemical zones trend in different directions, and are unaffected by twinning. In parts, sharp truncation of the chemical zones indicates resorption prior to crystal welding. Some of the welded-on crystals are twinned as well as zoned (see Fig. 7N); the twinning also developed prior to welding.

Twinning--Many of the andesine phenocrysts show polysynthetic twinning. The twin lamellae are fine and frequently irregular, resembling examples figured by EMMONS and GATES [1943]. Simple Carlsbad (see Fig. 7L) and complex interpenetration (see Fig. 6H) twins also occur. Combined forms consist of simple twins with either one (see Fig. 5C) or both of the twin halves showing polysynthetic twinning. Zoned crystals are twinned as frequently as non-zoned crystals (see Fig. 5, 6, and 7).

Some of the twinned-zoned crystals show primary twinning. Under the control of internal factors (low index and small obliquity) they grew spontaneously with the twin lattices in inverse positions (see Fig. 5C and 6E).

The fine twin lamellae of the polysynthetically twinned crystals are secondary (mechanical) twins produced by translation gliding due to external factors, for example, thermodynamic such as stress and temperature rather than by internal forces and spontaneous twinning during growth

[DONNAY, 1943]. (Some of the fine-twin lamellae may have developed during the preparation of the thin-sections.) Displacements of the chemical zones by these fine lamellae (see Figs. 5B, 6F, and 7M) cannot be detected under high power magnifications of the petrological microscope, hence the displacements must be of the order of atomic dimensions only. In Figure 7M there is a major displacement along a cleavage plane resulting from other stronger stresses whereas the fine lamellae to the left show no visible displacement.

The mechanically formed twins developed during cooling of the andesitic lava. The crystals were subjected to stresses set up during flow of the viscous lava [EMMONS and GATES, 1943]. Fractured crystals were subjected to further stresses arising from rapid crystallization of the groundmass. Intermediate to mild stresses were sufficient only to rotate the various parts of the twins distances of atomic dimensions along the space lattice. More intense stresses, however, operated in parts of the cooling lava and fractured some of the phenocrysts (see Fig. 7O). The fractures were subsequently occupied by narrow veinlets of groundmass constituents. Minute twin lamellae developed parallel to the direction of the fracture in the crystal sketched in Figure 7O. A few of the broken crystals were partially rotated, others were displaced along major cleavage directions (see Fig. 7M).

The writer agrees with EMMONS and GATES [1943] that twinning in most crystals developed late in the crystallization history of the lavas. They, however, consider that a plagioclase phenocryst ordinarily suffers little twinning where occurring in a fine-grained groundmass [p. 288, 1943]. The Bogana andesites seem to be an exception to this generality because they contain many twinned plagioclase phenocrysts in a fine-grained groundmass.

Zoning--Owing to the fact that neither a Universal stage nor an oil-immersion lens were available, precise statements concerning the compositions of the chemical zones cannot be made herein. Chemical zoning in the Bogana andesites is particularly well-developed in the andesine phenocrysts (see Fig. 4, 5, 6, and 7). The zoning is evident from refractive index and extinction angle variations.

Among the various andesine phenocrysts in the hornblende andesites, there occur: (1) Zoned crystals not twinned (see Fig. 5A, 6G, 7I, 7J, and 7K); (2) twinned crystals not zoned (see Fig. 7L); (3) twinned-zoned crystals (see Fig. 5B-D, 6E, 6F, and 6H); and (4) non-twinned, non-zoned crystals.

Inhomogeneities are evident in the majority of the phenocrysts, although non-zoned crystals free of inclusions are more or less uniform in composition. The width and character of the chemical zones, coupled with the type of twinning, lead to many varieties of andesine crystals.

Combinations of fine lamellar twinning and closely spaced, narrow chemical zones at right angles to the twin lamellae occasionally result in a cross-hatched appearance (see Fig. 6F) under crossed nicols.

Some of the variations are apparent only, due to the directions and positions obtained by random sectioning; some crystals, for instance, may not have been sectioned through the cores.

The principal characteristic of the andesine crystals is their marked discontinuities and they resemble examples described by PHEMISTER [1934] and by LARSEN IRVING, GONYER, and LARSEN [1938]. PALIUC [1932], has also indicated wide differences in composition between adjacent zones, and a great variability in the zonal succession.

All of the variations listed herein for the andesine phenocrysts of the Bogana hornblende andesites can be recognized in an area of two thin-sections, each approximately 3/4 inches × 3/4 inches. Normal, simple reverse, oscillatory, and oscillatory reverse zoning occur in the phenocrysts. In them, the type of oscillation varies within moderate limits. The smaller, groundmass feldspars are mainly non-zoned, but a few show normal zoning.

The variable character of the zoning in the phenocryst feldspars, is a direct manifestation of their complicated growth history. The principal trend in the zoning is less calcic from the core to the periphery.

Normal zoning--Both (a) gradational and (b) discontinuous normal zoning occur in a small number of the phenocrysts.

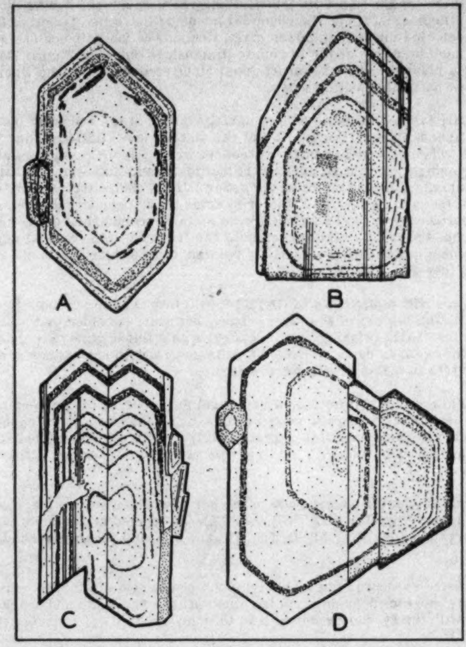

Fig. 5--Sketches of plagioclase phenocrysts as observed in polarized light; (A) zoned, non-twinned phenocryst with welded-on border crystal on left (× 67); darker zone represents a band of incompletely resorbed colorless inclusions, confined to one chemical zone; core and most zones extinguish at seven degrees to the long axis of the crystal, alternating zones at two degrees; (B) twinned phenocryst with oscillatory reverse zoning (× 63); core extinguishes at 24°, surrounding zone at 32°, remainder alternating at 24° and 32°; the fine twin lamellae do not appear to have displaced any chemical zones, but extinction positions in any one zone are reversed on either side of the minor glide planes; (C) primarily twinned phenocryst with oscillator reverse zoning (×33); chemical zones nes appear in inverse positions on either side of the twin plane, and, like the terminal crystal faces, have re-entrant angles; left-hand twin with fine lamellar twinning about parallel with the chemical zones; the core and similarly shaded zones extinguish at three degrees to the lamellae, other zones at six degrees; small welded-on border crystals are independently twinned and zoned; (D) zoned phenocryst showing primary twinning displacing the chemical zones and crystal edges (× 73); in right-hand twin portion, the core extinguishes at 14° to the twin plane, the next zone at 22° and the outer zones at 14°; in the central twin portion, the core extinguishes at 12° to the twin plane, the next zone at 20°, and the outer zones at 12°; in the left-hand twin portion, the core extinguishes at four degrees to the twin plane the next zone at six degrees, the remaining three zones alternating at four degrees, six degrees, and four degrees; the small border crystal on the left-hand side of the phenocryst has been welded-on.

(a) Those with normal gradational zoning have no clearly separated zones. They show unidirectional, undulose extinction, indicating a gradual and continuous change from a more calcic core to a more sodic periphery (see Fig. 7J; darker stippling indicates positions of extinction of

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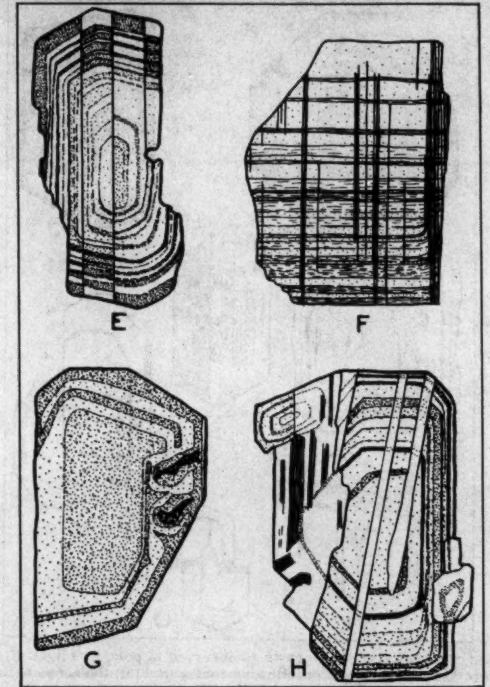

Fig. 6--Sketches of plagioclase phenocrysts as observed in polarized light; (E) complexly twinned and zoned phenocryst showing partial corrosion of outer edges (×50); in the second twin-portion from the left-hand side of the crystal, the core and similarly shaded zones extinguish at 20°, alternating zones at 28° to the twin plane; there is a difference of only 1/2 degree to one degree in the extinction angles between the outer and inner zones of each alternating set; the twin portions on either side of the second twin-portion from the left have the core extinguishing at 0°, in the same position as the darker shaded zones; lighter shaded zones extinguish at 15 °; zones in the positions of extinction in one twin-portion, show polarization colors on the opposite side of the twin plane (zoning is oscillatory in one twin-portion, oscillatory reverse in another twin-portion); (F) phenocryst with closely spaced chemical zones at right angles to fine twin lamellae, producing pseudo-cross-hatching (× 33); shows oscillatory zoning, the chemical zones extinguishing alternately at 11° and six degrees; the twin lamellae do not apparently displace the chemical zones, but in between the very closely spaced twin lamellae, the chemical zones are scarcely distinguishable; (G) phenocryst with oscillatory reverse zoning; the darker shaded portions indicate more sodic zones in the position of extinction (× 50); lighter shaded areas extinguish at nine degrees; the chemical banding is interrupted in the outer zones on the right-hand side by indefinite, partially resorbed inclusions (black areas); (H) twinned and zoned phenocryst with several welded-on border crystals (× 37); left-hand twin half with narrow polysynthetic twinning and one faintly-marked chemical zone near the center; right-hand twin with many chemical zones and broad lamellar twin bands trending obliquely; the core of this portion extinguishes at seven degrees to the twin planes, the zone next to it at 0°, the following zone at seven degrees, the next zone at 29°, and the remainder at seven degrees and four degrees, alternating

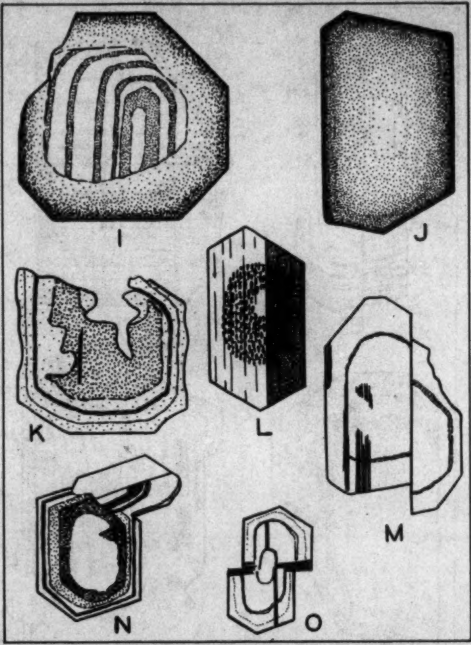

Fig. 7--Sketches of plagioclase phenocrysts as observed in polarized light; (I) phenocryst with rounded core showing normal discontinuous zoning (× 113); the zones of the core are sharply truncated by corrosion, and the remnant crystal is surrounded by a jacket with continuous zoning, becoming more sodic outwards; (J) non-twinned phenocryst with continuous zoning (× 230); the darker shaded portions represent the more sodic zones; (K) corroded, homogeneous core of calcic plagioclase, enveloped in a corroded but more sodic jacket showing faintly-marked oscillatory zoning (× 90); the dark elongated area in the core represents a needle-like crystal of hornblende; (L) simple twin of andesine (extinction angle 16 degrees) showing central zone of inclusions but no chemical zones (× 60); the inclusions tend to lie along cleavage directions (vertical lines in sketch) and appear to be unaffected by the twinning; (M) phenocryst in which non-rotational slip has occurred along a cleavage plane (× 72); shows offsetting of the crystal contour and chemical zone; lamellar twinning is confined to the left-hand portion of the crystal; (N) welded-on crystals (× 55); the top crystal is a simple twin with only the lower twin half exhibiting chemical zoning; the zoning is comparable in character to that of the lower crystal; the lower crystal possesses a corroded core, and in the surrounding chemical zones the outward change is towards more calcic plagioclase, so that the zoning is oscillatory reverse; (O) crystal offset across the center by partial rotational slip (× 67); the chemical zones have been laterally displaced, and secondary twin lamellae have been developed parallel with the direction of displacement

more sodic outer zones). Similar normal zoning is shown in the individual halves of twinned crystals, in crystals with inclusions in the core, and in crystals with peripheral zones of inclusions.

(b) Those with normal discontinuous zoning also show a variation of calcic to sodic plagioclase from within outwards. Since the changes were obviously discontinuous, crystallization of these phenocrysts occurred in marked stages.

In this group were developed several types of phenocrysts, such as crystals with (1) the core enveloped in an outer jacket of marked compositional difference, the jacket showing normal discontinuous zoning in itself; (2) both core and outer jacket separately showing gradational zoning, as evident from the undulose extinction in each portion; (3) the core and the jacketing zone each showing discontinuous zoning; and (4) the core and the jacket of different extinction, neither being zoned in themselves, and each therefore of uniform but different chemical composition.

Simple reverse zoning--Sharply marked central zones in certain of the phenocrysts of andesine, are less calcic than the peripheral zones, thus resulting in simple reverse zoning. Several combinations are recognised in this group; for example, (1) core becoming less calcic from within outwards,(that is, shows normal discontinuous zoning), but the surrounding jacket becoming more calcic from within outwards in discontinuous stages; (2) a sharply marked, rounded core with reverse zoning, surrounded by an idiomorphic jacket of more calcic plagioclase showing normal gradational zoning (see Fig. 7I); and (3) discontinuous zones becoming more calcic from within outwards.

Other phenocrysts that come under the category of (2), possess cores with considerably embayed outlines. PHEMISTER [1934] regards phenocrysts like these as being, in a sense, xenocrystic and representing early-formed crystals in a magma cognate with the rock now enclosing them.

Oscillatory zoning--Several phenocrysts show more complex zoning and have larger numbers of sharply marked zones. One set of zones with similar compositions alternates with a second set of zones of similar compositions. The compositions of the two sets differ however. One set is less calcic than the core and the other set, giving rise to normal oscillatory zoning (alternating or rhythmic zoning of PHEMISTER [1934]. Each zone shows a slight, progressive outward change from more to less calcic plagioclase. The various growth shells oscillate from more or less calcic, but become more sodic towards the periphery.

Oscillatory reverse zoning--In a few phenocrysts, less calcic cores are enveloped in more calcic outer jackets each having more calcic peripheral zones (see Fig. 6G). The zones are discontinuous and oscillating, but in the reverse order. Some phenocrysts with oscillatory reverse zoning, are enveloped by a peripheral jacket of composition analogous to that of the inner core. Others have a basic (An-rich) core, enveloped by an even more basic zone, which in turn is jacketed with four successive zones of the two types alternating.

Most of the individual zones in the oscillatory reverse group, extinguish as single units, but a few show undulose extinction, indicating gradational chemical changes within them. The majority of the oscillatory zoned crystals are idiomorphic, and the zones usually (see Fig. 4) exhibit much the same idiomorphic character as the crystal. The chemical zones are generally narrow on the sides of the crystals, and broad in the direction of the c-axis, where growth was apparently more rapid.

One of the most remarkably zoned andesine phenocrysts in the Bogana andesites, shows upwards of 45 well-marked, idiomorphic chemical zones (see Fig. 4). The crystal is only two mm long. The central core extinguishes at 44° to the prominent cleavage directions. Birefringent inclusions of incompletely resorbed plagioclase in the core extinguish together at 22°. The core is surrounded by two main jackets, each less calcic than the core and each with several oscillatory growth zones. The outer of the two jackets is less calcic than the inner. The zones of the inner jacket extinguish at 27° nearest the core, and are enveloped by oscillatory reverse zones extinguishing variously at nine degrees, 14°, and 18°, thus indicating increased An-content from within outwards. The euhedral character of the core is very pronounced and earlier-formed zones closely follow the outline of the core. Later-formed zones of the inner jacket have developed in such a way as to eliminate one of the original crystal faces (see zones above top right-hand corner of the core). The zones of the outer jacket grew parallel with those of the outer portion of the inner jacket.

Distribution of zoned-twinned plagioclase crystals

Plagioclase feldspars with well-marked combined zoning and twinning structures are abundant and well-developed in the Bogana lavas on Bougainville Island. Comparable types occur in hornblende andesite from Diit village on Tacloban, Leyte in the Philippine Islands, although the feldspars are more metasomatically altered. Similar feldspars occur in hornblende and augite hornblende andesites in the Aitape district, north coast of New Guinea, and in hypersthene and hornblende andesites (from river-worn pebbles) near the Dobodura airstrip, southeast Papua. The andesites from Urasi Island in the Amphlett Group contain similar twinned-zoned feldspars. In dacite from southwest of Bolu Bolu on Goodenough Island, D'Entrecasteaux Group, some of the plagioclase feldspars are likewise complexly twinned and zoned [BAKER and COULSON, 1948].

HOMMA [1936a] has illustrated similar twinned-zoned plagioclase from the dacite of Macyama, Unzen, Japan, from biotite hornblende andesite at Simpora Peninsula, Northern Borneo, and from the two-pyroxene andesite of Shin-take, Kuchierabu, southwest Japan. Well-developed oscillatory zoned plagioclase crystals have also been recorded by FENNER [1926] in two-pyroxene andesite from the Katmai magmatic province in Alaska.

Twinned-zoned feldspars of the kind described from the Bogana andesitic lavas, are thus widespread in the intermediate lavas forming the larger part of the Quaternary volcanic girdle that encompasses the Pacific Ocean. Such types, however, are not confined to this geologically young series of lavas, as HILLS [1932] has recorded crystals of plagioclase with oscillatory zoning and complex twinning in the same crystal from much older lavas, namely from meta-andesite in the Marysville igneous complex, Victoria, where the lavas are of Upper Devonian age. Nor are they confined among Tertiary lavas to the Pacific girdle for PALIUC [1932] has described similar occurrences from dacites and andesites at Baia Mare (eastern Carpathians) and in the Transylvanian Alps.

Petrogenesis

The association of many types of zoned (inhomogeneous) and non-zoned (homogeneous) andesine phenocrysts in the Mt. Bogana andesites, indicates a complex crystallization history that no one simple scheme explains.

The homogeneous crystals were formed by slow crystallization conducive to complete diffusion and the attainment of equilibrium. Alternatively, conditions favored rapid diffusion that sometimes occurs in calcic lavas. The experimental results set out by WAHL [1946] indicate that where thermal diffusion-convection processes operate, inhomogeneities may not develop in all parts of an originally homogeneous magma. Hence it is possible for non-zoned andesine phenocrysts to form in the same magma chamber where several kinds of zoned andesine phenocrysts crystallize.

The inhomogeneous crystals require variable conditions for their development. In parts of the magma, conditions suited progressive crystallization. Consequently, normal gradational zoning occurred as evidenced by the presence in the lava of phenocrysts showing gradual transition from calcic cores to sodic borders.

In other parts of the magma, conditions favored rapid changes of temperature or periodic changes in the composition relationships between magma and phenocrysts. As a result, discontinuous crystallization of variously alternating calcic and sodic zones, produced normal discontinuous, simple reverse, oscillatory, and oscillatory reverse zoning. The narrowness and multiplicity of chemical zones in some phenocrysts, indicate that certain portions of the magma were more suited than others for the repeated formation of various zones.

Reasons for the discontinuous crystallization of growth zones have already been suggested by BOWEN [1913 and 1929], HOMMA [1932], PHEMISTER [1934], HILLS [1936], and LARSEN, IRVING, GONYER, and LARSEN [1938]. The diffusion-saturation theory advocated by HILLS [1936] and supported in its entirety by LARSEN, IRVING, GONYER, and LARSEN [1938], accounts for most discontinuities evident in the cores and inner zones of the Mt. Bogana andesine phenocrysts. It applies in particular to the rhythmical precipitation of the alternating shells constituting oscillatory zoned crystals. Slight irregularities in contour of some of the inner zones, were caused either by local impoverishment of mineralizers, thereby temporarily upsetting conditions of equilibrium and leading to partial resorption, or by movement of phenocrysts in convection currents [HOMMA, 1932] to positions where temporarily out of equilibrium with the magma.

Certain incongruities that exist between the compositions of inner cores and outer jackets, and the variability in composition of the outer jackets themselves, are thought to have been produced in parts of the magma remote from positions where cores and inner zones were formed. The outer jackets were evidently formed late in the crystallization history of the Mt. Bogana andesites, and not very long before lava extrusion.

In the volcanic vents and cupola magma, there existed considerable variations in conditions of equilibrium between magma and partially grown zoned phenocrysts. The variations were caused by frequent lava pulsations, brought about by discharge of mineralizers during periods uninterrupted by more violent volcanic activity, when new supplies of volatiles were rising regularly into the cupolas and vents. Such variations are indicated to some extent by the relationship between augite and hornblende in the andesites. The hornblende with its reaction zones and the augite with its two generations of crystals (phenocrysts and groundmass prisms), indicate oscillations of "wet" and "dry" conditions of crystallization [BAKER, 1946, p. 30].

Temperature and compositional changes accompanying resurgence of volatiles, were of sufficient frequency to influence the character of growth zones in the outer jackets of some of the andesine phenocrysts. Concomitantly, changes of pressure, associated with viscosity variations, especially in the vent regions, further complicated the conditions, and considerable turbulence would exist.

The operation of these processes in the vent regions, accounts for many of the anomalously zoned phenocrysts, for instance, non-zoned cores surrounded by narrow outer jackets with oscillatory reverse zoning.

The ultimate uprise of different fractions of the magma into cupola and vent regions, had brought variously zoned and non-zoned phenocrysts into positions where temperature gradients were either not established or only transient phenomena. In these positions, the requisite conditions existed for discontinuous zonal growth, hence the further development of oscillatory and oscillatory reverse zoning found in some of the outer jackets.

Thorough mixing of fractions of magma in which had formed the non-zoned and several types of zoned phenocrysts, would have to occur in the vent regions, in order to account for such a mixed assemblage of andesine phenocrysts so evident even in areas as small as the size of ordinary thin sections. Concurrently, all vestiges of schlieren that might be expected by such mixing of magma fractions, had to be smoothed out and lost during the mixing process. Turbulence became an important factor during this process, and the magma was probably too fluid for schlieren to be formed.

During pulsatory movements of magma in the vents, small border crystals of andesine became attached to some of the larger andesine phenocrysts. They were subsequently welded-in when a few additional narrow outer zones grew on the already well-formed phenocrysts.

Acknowledgments

The rocks studied from the Mt. Bogana-Torokina River area were collected by H. K. Werner in 1945, from an area covering approximately ten sq mi and extending from Empress Augusta Bay to half way up the slopes of Mt. Bogana. The photograph of Mt. Bogana (see Fig. 1) was supplied by the Department of the Army (Australia). The author is indebted to E. S. Hills of Melbourne University for suggestions and comments rendered during the work.

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(Communicated manuscript received August 6, 1947;
open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

THE ROLE OF HYDROMETEOROLOGY IN PLANNING THE WATER ECONOMY OF THE WEST

Merrill Bernard

About five years ago the writer had occasion to report upon the development of hydrometeorology [BERNARD, 1944]. Briefly summed up to that time, the story of this hybrid science was one of a rapid evolution, to which primary impetus had been given by the Corps of Engineers, then assuming heavy responsibilities in the Nation's flood-control program. Methods of determining flood-peak discharge had progressed from a function of the square root of the basin area, through statistical expressions of transposed flood experiences and empirical rainfall-runoff relations, to the transposition of the isohyetal pattern of an actual storm to and over a problem basin converting its runoff to stream flow by unit hydrograph or other flow-routing methods. Here unexpected questions were posed: "Could the transposed storm have been greater where it occurred?" "Is the transposition meteorologically justified?" In characteristic fashion the Army Engineers, without hestiation, embraced a heretofore unutilized science, meteorology, and combined it with one already established in the field of civil engineering, hydrology, to create hydrometeorology. Since then the original relationship has become somewhat obscured, and hydrometeorology is now best defined as the dynamic meteorology of storms and the hydrology of storm rainfall and snowmelt with particular attention to their distribution through time and over area.

It should be noted that up to the present hydrometeorology has found its principal application in the determination of the physical upper limit of storm rainfall. The explanation is, that the United States Corps of Engineers, while not necessarily using the maximum possible values of storm rainfall as the basis of design, have felt that such knowledge should be made available to the designer of any important flood-control structure. The assignment of frequency to storm magnitude must await the final processing of the data for approximately 2000 flood-producing storms of record, a continuing project in certain of the offices of the Corps of Engineers.

The Hydrometeorological Section of the United States Weather Bureau has been financed, since its beginning in 1937, by the Corps of Engineers and devotes itself exclusively to the Corps' problems in collaboration with the field offices of the Corps of Engineers. More recently the Cooperative Studies Section has been established in the United States Weather Bureau to serve the United States Bureau of Reclamation in similar fashion, working directly with its Hydrologic Section of the Branch of Project Planning at Denver.

The earlier hydrometeorological studies dealt with the Central Plains and regions to the East. In this portion of the country storm transposition could be taken literally for the reason that rainfall was induced primarily by frontal action, or the intensive convection of thunderstorms, orography being a lesser influence in all except the more mountainous parts of the Eastern States. In recent years the programs of the Corps of Engineers and the Bureau of Reclamation have taken hydrometeorology into the West. Here a reversal of emphasis has been introduced in that extensive mountain ranges impose the dominating influence on the mechanism of storm rainfall.

This change of scene has called for a restatement of concept and the redefinition of storm transposition. It is simple enough to transpose a storm within large areas of meteorological homogeneity such as the Central Plains under the assumption that it could, with equal probability, have occurred over the basin being considered. In the West, where orographic influences dominate, it becomes obvious that transposition, as previously defined, loses all reality and that a storm experienced in one mountainous locality cannot occur in another without profound modification of airmass properties, dynamics of mechanism, magnitude of rainfall, and configuration of isohyetal pattern. Storm transposition, therefore, is giving way to storm synthesis, an intimate analytical knowledge of the storms within a region making it possible, literally, to disassemble then where they occurred and reassemble them within the environment of the region in which the problem basin lies.

While the orographic control characteristic of western regions has introduced an element of stability into storm analysis, the advantage is offset in part by the fewer data available to the analyst.

It is not unusual to find problem basins in which no observational data exist. This limitation has quickened the ingenuity of the hydrometeorologist and has forced the development of methods of indirect analysis which do him credit.

Thus we now find the hydrometeorologist probing beyond the covers of meteorological textbooks on hydrodynamics, aerodynamics, and thermodynamics, his experience having proven that uniform flow, homogeneous mass, simplified structural models, and uniform distributions are oversimplifications of storm situations in which elevation, ground slope, aspect, turbulent eddy action, and other factors, cannot be disregarded.

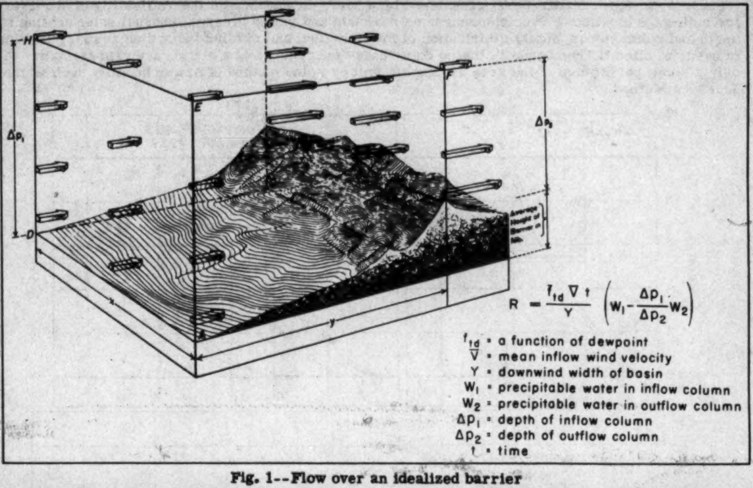
The ultimate aim, of course, is to evolve an orderly equation expressing rainfall as an explicit function of all variables. Theoretical meteorology has made remarkable progress in recent years, yet only a few of the variables can be handled with facility, and their interrelations are still not well known. It can be seen that the statistical approach, with unlimited data, would produce a rainfall equation capable of reproducing rainfall amounts within the errors defined by the known variables. However, lacking full understanding of the physical processes and complete observational coverage, the present approach is to join sound meteorological reasoning with statistical analysis. Physical considerations prescribe the pattern of the storm model in terms of an equation of continuity. In this form the pitfall of unrelated occurrences masking the real variables is avoided. With the model and formula thus physically determined, statistical analysis is employed to check the formula, to identify the variables as far as possible and to determine coefficients and proportionality constants. It will be interesting to see how this approach has been utilized in the development of rainfall equations applicable to West Coast regions, and to the evolution of rainfall models.

Anyone who has examined topographical and isohyetal maps in the Western States must have recognized the close relationship between them. The connection is most apparent in mountainous regions near the coast where onshore winds have an undepleted moisture charge and offshore winds are dry, having passed over high, rugged inland areas.

Several years ago, the Hydrometeorological Section was assigned the problem of determining the maximum possible precipitation over the Sacramento Basin. Drawing on its experience in dealing with similar problems in eastern United States, the Section proceeded to adapt its already-developed relationships to the Sacramento Valley. The theoretical flow model simulated that of a series of thunderstorm cells, that is, one in which each air column crossing the basin would be forced to release precipitation equivalent to a certain amount of convective overturning. In the adaptation of this model to the Sacramento, the moisture charge of the air entering the cell was assumed to be depleted by the inflow barrier of the Coast Range. A further assumption was that the height of the cell would vary directly with sea-level dewpoint. Observed rainfall measurements and meteorological data associated with them were used to check the accuracy of the rainfall equation. Finally, a statistical analysis of the index parameters produced critical values which, when used in the equation, yielded the estimate of maximum possible precipitation.

In the next report, which dealt with the Los Angeles area, what was felt to be a more realistic flow model was adopted. Fundamentally, it was assumed that the maximum possible precipitation in this region would be predominately orographic in nature and that any nonorographic contribution would be taken into account in the empirical phases of the investigation. The model contained a "nodal surface," which is defined as that above which orographic influences are assumed to be absent (see Fig. 1). As before, theoretical treatment of the continuity principle prescribed the form of the rainfall equation, and statistical analysis of observed data provided the means of writing the equation in such determinable variables as geometric basin properties, surface dewpoint, and sea-level geostrophic wind. Insertion of observed values of the parameters into the formula reproduced the observed rainfall values over 10,000 sq mi with an unusual degree of accuracy (see Fig. 2). An empirically determined critical combination of values, when introduced into the equation, yielded the estimate of maximum possible precipitation.

In a recent major basin report, dealing with the San Joaquin Valley, further refinements to the theory were introduced. Physical considerations led to the conclusion that the nodal surface should be higher over the Sierra than over the Coast Range, and the equation was accordingly modified (see Fig. 3). Also taken into account was the spillover effect, which deals with rain formed in the air above one basin but which falls in another basin downwind. Theoretical hydrodynamics, in demonstrating that air must begin to rise before it reaches a mountain barrier, leads to the conclusion that orographic rain can fall in advance of a mountain range. Lacking the means of dealing quantitatively with this upwind effect in a theoretical way, statistical treatment of rainfall observations at both mountain and valley stations yielded results that were consistent and meteorologically


H
G
F
E
D
A
Δp_1
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x
y
Average height of barrier in mb.
$R = \frac{\bar{f}_{td}\,\bar{V}\,t}{Y}\left(W_1 - \frac{\Delta p_1}{\Delta p_2} W_2\right)$
$\bar{f}_{td}$ = a function of dewpoint
$\bar{V}$ = mean inflow wind velocity
Y = downwind width of basin
W_1 = precipitable water in inflow column
W_2 = precipitable water in outflow column
Δp_1 = depth of inflow column
Δp_2 = depth of outflow column
t = time

Fig. 1--Flow over an idealized barrier

acceptable. A breakdown of total storm rainfall into inflow spillover, upwind, orographic, and outflow spillover rainfall, is illustrated in Figure 4. The importance of the upwind effect was brought out when it was found that nearly 20 per cent of the orographically-produced rain fell in advance of the mountains in those storms subjected to analysis. The spillover effect, relatively unimportant for large basins, was determined to be highly significant for small basins, especially so when the ground slope was steep.

In the evolution of the West Coast rainfall model, successive refinements have been introduced as theory that understanding has advanced. Paucity of observational data has been a decided handicap, yet it has been possible to compute rainfall values which compare very reasonably with those observed. The greatest successes, however, have been attained in computations for relatively large basins. The degree of security of these total basin values will not be reached in the case of the smaller sub-basins until many more observational data are available and after certain theoretical investigations are completed. It is encouraging to know that theoretical knowledge is accumulating, and it is hoped that the observational program will keep pace.

Difficulties encountered elsewhere become especially aggravated in the interior mountain regions. They arise from the general inadequacy of basic hydrologic and meteorological data, the lack of facilities for processing these data to increase their utility, and the excessive orographical influence on storms occurring within these regions.

Methods for making reliable estimates of the water volume of seasonal snowfall are subject to considerable improvement although some progress is being made in this direction. Here the lack of data on water-equivalent depth is overshadowed by the almost complete absence of information leading to the identification of runoff-producing residual areas as they change their size, shape, and relative positions throughout the melting period.

The determination of maximum possible storm rainfall in inland mountain regions must, for the present, be limited to the modification and adjustment of those storms of record which have occurred on the problem basin. The results are not satisfying since the greater number of basins have experienced only one or two major storms within the period of record, and in some there have been no major storms recorded.

The inferiority of hydrological and meteorological data in mountainous regions does not result from poor observational techniques, which are just as good in this region as elsewhere, but rather from lack of representativeness, occasioned by a heavy dependence on the availability of observers for rain-gage locations. Such phenomena as mountain and valley breezes, unequal solar heating of north and south slopes, spotty distribution of precipitation, and similar factors so restrict the size of areas of climatic homogeneity that in many cases the records at a station are representative of only a small percentage of the area assignable to it by some method of proportionality such as the Thiessen Method.

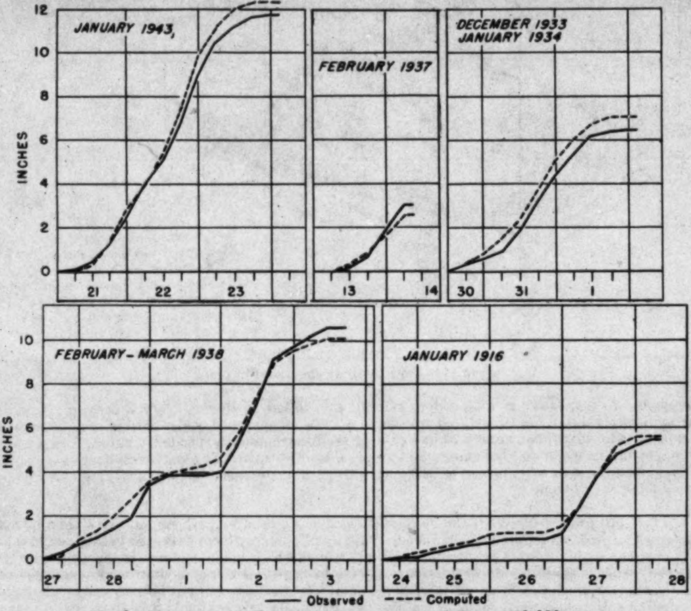
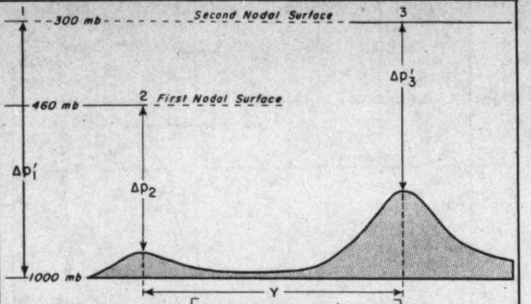


Fig. 2--Mass curves of maximum average rainfall over 10,000 sq mi

Obviously, the situation can be improved by increasing the density of the observational network and by selecting station sites with due regard for representativeness. Unfortunately, the number of stations is definitely limited by the availability of funds while the ease of maintenance is often the deciding factor in the choice of station sites.

To offset these discouragements, it has been found that the available data can be processed in such a way that their representativeness is materially improved. This has been accomplished by giving expression to such physiographic parameters as elevation, slope, aspect, and environment. The double mass-curve method for adjusting the individual precipitation record to the apparent climatic regime of its locality is well known. This procedure attains its maximum usefulness in the mountain regions, where changes in station location of a few miles may result in significant differences in elevation, slope, aspect, and exposure. A recent study assigned to the Cooperative Studies Section by the Bureau of Reclamation has been successful in developing from available data reliable average annual and seasonal precipitation maps for western Colorado, where the topography is probably as rugged as anywhere in the inter-mountain region and where the type, distribution, and location of observation stations are typical of the mountainous West. This statistical analysis of the relation of precipitation to topography revealed five topographic parameters to which expression could be given (see Fig. 5).



$$R = \frac{f_{td}\,\overline{V}\,t}{Y}\left[\left(W'_1 - \frac{\Delta p'_1}{\Delta p'_3}W'_3\right) - \left(W_1 - \frac{\Delta p_1}{\Delta p_2}W_2\right)\right]$$

f_{td} = a function of dewpoint
$\overline{V}$ = mean inflow wind velocity
Y = downwind width of basin
W_1 = precipitable water in inflow column
W_2 = precipitable water in outflow column
Δp_1 = depth of inflow column
Δp_2 = depth of outflow column
t = time
W'_1 and W'_3 = precipitable water in columns at inflow and second barrier, respectively, as measured to height of second nodal surface
$\Delta p'_1$ and $\Delta p'_3$ = depths of columns from ground surface to second nodal surface at inflow and second barrier

Fig. 3--Schematic orographic model, San Joaquin River Basin

The relationships established for October-April precipitation are indicated in the figure. Using these relationships, it was possible to compute precipitation at the intersection points of a coordinate grid over the area. From these computed values an isohyetal map of normal October-April precipitation was constructed. A map of annual precipitation was developed in a similar manner. The topographic parameters responding to the correlation accounted for about 85 per cent of the original variation in precipitation with correlation coefficients over 0.90 and with standard errors less than three inches in both cases. Although elevation accounts for most of the variation, the correlation was definitely improved by the introduction of the other four parameters, namely, rise, exposure, orientation, and zone of environment (see Fig. 6).

The contrast between the mean annual precipitation map constructed by this method and that drawn in the usual way is striking (see Fig. 7). The statistically constructed map shows isohyets for values up to 60 inches whereas the highest isohyetal values on the published isohyetal map of the area is 30 inches. A precipitation record for a recently installed station at Wolf Creek Pass, not included in the present study or in the preparation of the earlier map, confirms the higher values. An eight-year average annual depth of 48.39 inches for this station is believed to be reasonable, and in conformity with values indicated by vegetal cover and other physiographic evidences of precipitation variability.

Crusaders for water conservation are legion. They range from the powerful Federal agencies engaged in far-reaching programs of area development to the water master who conscientiously distributes among his water customers the infinitesimal part of the season's water supply allocated to him. While it is true that there is a long experience in the management of western water supplies, of which the managers may be justifiably proud, to the more observant there is some feature lacking in the plans of the future for an economy so heavily dependent on water. The inventory does not include, nor does the system of bookkeeping appear to provide for the source of the raw material, the genesis of all terrestrial water, precipitation. It seems conservative to say that there does not now exist a source of quantitative precipitation data which we must have as the starting point in water accountability in terms sufficiently refined to meet the needs of the analyst, and with sufficiently detailed coverage to serve all interests engaged in the development of western water resources.

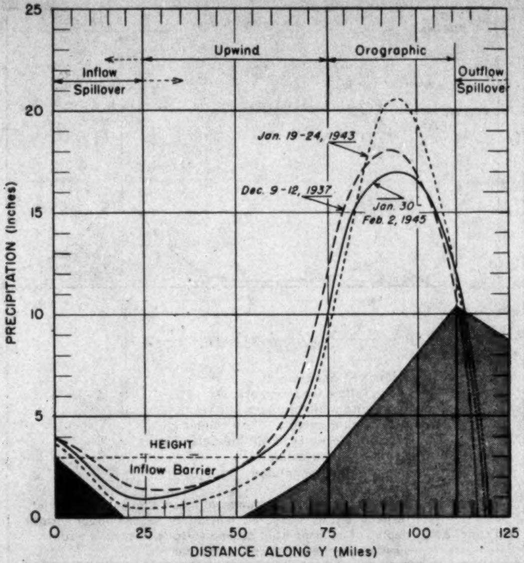


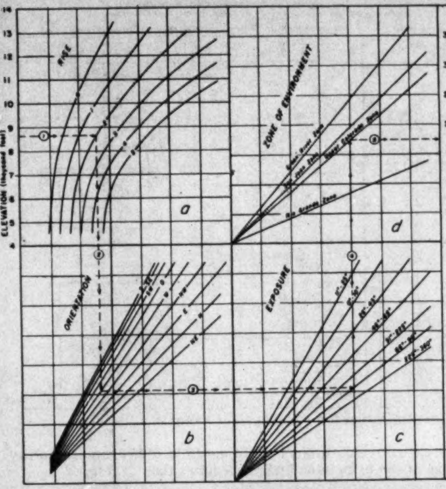
Fig. 4--Storm profiles, San Joaquin River Basin

Here the writer expects to be challenged. His attention will be called to the money and effort spent by at least a dozen Federal agencies and as many state offices in measuring precipitation. He will be reminded of the difficulties of gathering such data in the remote regions, of the unrepresentativeness of the longer precipitation records which are found at low or intermediate elevation; of the extensive streamflow observations that presumably measure directly all the water in which we are interested; of the remote relation of ground water to precipitation. These objections have the guise of reality. They do not, however, represent obstacles which are insurmountable.

It is not too great a strain on the imagination to conceive of maps of the kind shown portraying precipitation as a function of detailed geographical distribution. These maps, made to represent variations in duration from annual through storm-period precipitation, as well as frequency of occurrence, would identify the areas of water abundance and deficiency and the gradations lying between these extremes. The profound effect on planning of the availability of such maps needs no emphasis.

Succeeding the maps of precipitation would be similar ones showing the deductions to be made for such unpreventable losses to water supply, as evaporation and transpiration. As these phenomena are hydrometeorological in nature, the hydrometeorologist is prepared to engage in the research required to successfully measure and ultimately to compute them in terms of the characteristics of air masses passing over a region.

Maps portraying infiltration characteristics would lead to the identification of those areas whose water-producing potentialities justify, in varying degree, the establishment of stream-discharge measurements in preparation for ultimate exploitation. And, finally, the map expressing the integration of the precipitation, evapotranspiration, infiltration, surface runoff, and geological maps would reveal for the first time the connecting links relating ground water to its initial source, precipitation.



Legend

Elevation--Station elevation, expressed in ft above mean sea level

Rise--Difference in elevation between the station and the highest point within a five-mile radius, expressed in thousands of ft

Orientation--Direction of the exposure which is defined above, expressed to eight points of the compass (in cases of equal exposure in two directions, orientation was taken as that of the exposure which opens downslope, that is, toward lower elevation)

Exposure--The sum of those sectors of a 20-mile radius circle, centered at the station, not containing a barrier 1000 feet or more above the station elevation, expressed in degrees of azimuth

Zone of environment--Includes those areas having the same general direction of moisture inflow and similar relation to the principal mountain ranges

Fig. 5--Relation between normal October-April precipitation and topographic parameters for western Colorado based on period, 1910-1945

Thus, a future is envisaged in which the meteorologist, the hydrologist, and the ground-water geologist, working together, would create a planning base of factural knowledge that, in a very literal sense, would become the keystone of western water conservation.

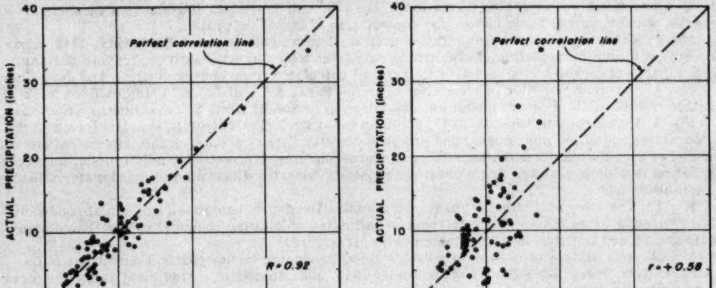


Fig. 6--Comparison of computed October-April precipitation based on period 1910-1945 (a) precipitation computed from graphical relation with all topographic parameters; (b) precipitation computed from linear relation with elevation alone

The following is a selected list of reports prepared by the Hydrometeorological Section of the United States Weather Bureau in cooperation with, and under flood-control funds provided by the Corps of Engineers, Department of the Army. Selection of a report for inclusion in the list is

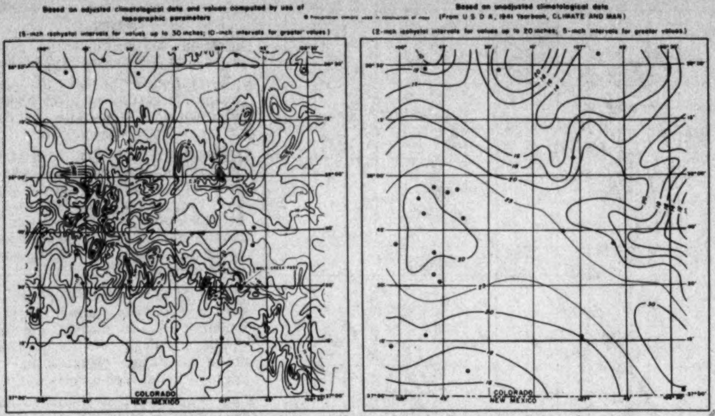

Based on adjusted climatological data and values computed by use of topographic parameters
(5-inch isohyetal intervals for values up to 30 inches; 10-inch intervals for greater values)
● Precipitation stations used in construction of maps
Based on unadjusted climatological data
(From U S D A, 1941 Yearbook, CLIMATE AND MAN)
(2-inch isohyetal intervals for values up to 20 inches; 5-inch intervals for greater values)
COLORADO
NEW MEXICO
COLORADO
NEW MEXICO

Fig. 7--Normal annual precipitation for southwestern Colorado

based upon the report's contribution to progressive developments in the field of hydrometeorology. Unlisted reports involve local applications of the principles and procedures thus developed.

No. 1, Maximum possible precipitation over the Ompompanoosuc Basin above Union Village, Vermont, 1943, introduces the concept of precipitable water (moisture charge) as a function of 1000-mb dewpoint, and adds to storm-transposition procedure the adjustment of major-storm rainfall by extrapolation from observed to maximum possible moisture charge.

No. 2, Maximum possible precipitation over the Ohio River Basin above Pittsburgh, Pennsylvania, 1942, modifies storm adjustment by introducing a convective-rainfall storm model, yielding effective precipitable water, also a function of the 1000-mb dewpoint. Introduces a theoretical formula, with empirical basin factor, for computation of snow-melt rates.

No. 3, Maximum possible precipitation over the Sacramento Basin of California, 1943, modifies storm model for depletion of effective precipitable water by windward orographic barrier, and compares observed basin rainfall with rainfall calculated from observed winds and dewpoints.

No. 4, Maximum possible precipitation over the Panama Canal Basin, 1943, involves intensity-duration modification of storm model for adaptation to region of limited convection in major storms.

No. 5, Thunderstorm rainfall, 1947, contains exhaustive climatology of thunderstorms in the United States and allied phenomena, and analysis of reliability of areal-rainfall determinations as function of gage distribution and density. Introduces modified convective-rainfall model for use in estimating maximum possible from observed rainfall. Includes discussion of thunderstorm theory and a bibliography.

No. 18, Tentative estimates of maximum possible flood-producing meteorological conditions in the Columbia River Basin, 1945, introduces estimates of limiting seasonal rather than storm precipitation, and limiting long-duration snow-melt factors.

No. 20, An estimate of maximum possible flood-producing meteorological conditions in the Missouri River Basin above Garrison Dam site, 1945, adds sesonally varied maximum rainstorm to a progressive estimate of seasonal snow accumulation.

No. 21B, Revised report on maximum possible precipitation, Los Angeles area, California, 1945, develops realistic model for computation of orographic precipitation on windward slope.

No. 22, An estimate of maximum possible flood-producing meteorological conditions in the Missouri River Basin between Garrison and Fort Randall, 1946, adds seasonally varied maximum snow storm to types of estimates made in No. 20.

No. 23, Generalized estimates of maximum possible precipitation over the United States east of the 105th meridian, for areas of ten, 200, and 500 sq mi, 1947, includes detailed discussions of development of latest convective-rainfall storm model, storm-transposition practices, and other procedures involved in estimates of maximum possible rainfall.

No. 24, Maximum possible precipitation over the San Joaquin Basin, California, 1947, further modifies orographic-rainfall model for two-barrier basin, quantitative corrections being added for both leeward and windward spillover of rainfall and for upwind effect.

The following is a selected list of reports prepared by the Cooperative Studies (Reclamation) Section of the United States Weather Bureau. This Section, organized in December 1945, is maintained jointly by the Weather Bureau and the United States Bureau of Reclamation to provide hydrometeorological studies for the latter agency. Selection of a report for inclusion in the list is based upon the report's contribution to progressive development in the field of hydrometeorology. Unlisted reports involve local applications of the principles and procedures thus developed.

Report No. 2, maximum possible flood-producing meteorological conditions in the East Muddy Creek Basin above Spring Creek Dam site, Gunnison and Upper Colorado River Basin Studies, 1946, describes a new method for obtaining an index of the maximum seasonal snow accumulation by adjustment of all storms in wettest seasons to maximum moisture charge.

Report No. 3, Maximum possible flood-producing meteorological conditions in the Taylor Basin above Almont Dam site and Taylor Park Dam, Gunnison and Upper Colorado River Basin Studies, 1947, introduces a new procedure for estimating maximum snowmelt rates from critical temperature sequences and indices of the snow pack.

W. B. Tech. Paper No. 4, Topographically adjusted normal isohyetal maps for western Colorado 1947, statistical analysis of topographic influences on precipitation to provide more accurate mean precipitation maps in mountainous regions.

Note--These reports are not available for distribution but they may be inspected at the Weather Bureau Library in Washington, D. C., or at the field offices of the Corps of Engineers, Department of Army, and the Bureau of Reclamation.

Reference

BERNARD, MERRILL, The primary role of meteorology in flood flow estimating, Trans., Amer. Soc. Civ. Eng., v. 109, 1944.

Climatologic and Hydrologic Services,
U. S. Weather Bureau,
Washington 25, D. C.

(Manuscript received March 17, 1948; presented at the West Coast Regional Meeting, Berkeley, California, February 13, 1948; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

HOW LONG SHOULD EXPERIMENTAL WATERSHEDS BE CALIBRATED?

H. G. Wilm

Abstract--A statistical method for estimating the necessary duration of a watershed experiment is presented that applies especially to the type of study in which two or more watersheds are placed under control by means of stream-gaging stations and other equipment. After a suitable period of calibration, all but one may be subjected to some form of treatment such as a change in land use or the removal of forest cover. With minor modifications the method may be applied to other kinds of experiment in which regression analyses are used to obtain statistical control of the variables under scrutiny.

Equations are derived with which the investigator can estimate the number of observations which is likely to be required before and after treatment, in order to demonstrate that a difference or treatment effect of any chosen order of magnitude is larger than might have been expected by chance.

In recent years a number of investigations have been established, using small watersheds for testing the influences of vegetation or of watershed treatments upon water yields, flood peaks, and related variables. The organization of such experiments follows a fairly well established procedure: Two or more watersheds are placed under careful control by means of stream-gaging stations, sediment basins, rain gages, and other equipment for measuring the desired factors. Records are collected for a period of years, during which all of the controlled areas are maintained in their original condition without experimental treatments. This is conventionally called the calibration period; its purpose is to show the natural relation of the watersheds to one another, and to provide a means by which uncontrolled variation, due to climate and other factors not associated with the treatment, may be minimized by statistical methods. At the close of this period all but one of the areas may be subjected to treatment and further records are taken until it is felt that the effects of the treatment, if any, are satisfactorily demonstrated.

Even before the treatments are applied in this kind of study, an important question comes up to perplex the investigator: How long should the watersheds be calibrated? If too short a period is allowed, the experiment may lack satisfactory precision; it may not be possible to demonstrate the real nature of even fairly large effects of treatment. On the other hand, if the period is longer than necessary, the final answers are postponed and the total cost of the investigation becomes undesirably large. Watershed studies are expensive at best; and because every additional year of calibration means another heavy charge against the net value of experimental results, it is desirable to estimate as precisely as possible how long the periods before and after treatment must be in order to supply satisfactory results.

The purpose of this article is to show how the necessary minimum length of a calibration period and of a whole watershed experiment may be estimated by logical and statistical methods. The latter is an extension of procedures for analyzing the results of watershed studies which have been presented in an earlier article [WILM, 1944].

Preliminary considerations--The use of the method to be described is predicated on several fundamental assumptions. The first of these is that, within any period which is likely to include the duration of the study, the variation of water yields or other factors to be investigated may be taken as random and normally distributed. As borne out by other studies [HOYT and LANGBEIN, 1944] and the writer's own experience, ordinarily water yield data can be expected to conform to this assumption within practical limits. Actually it is probably not as important, however, as a second assumption: That, when a regression equation is calculated to show the relation of yields from one watershed to those from a second area, the residual deviations around the regression line are random and conform reasonably well to a normal frequency distribution.

One more requirement is fundamental to the success of the whole analytical method. Because errors and variation may be associated just as much with data from the control watershed, X, as with those from the area to be treated, Y, it is necessary that their yields be correlated to a relatively high degree [EISENHART, 1939]. Otherwise enough bias may be introduced into the regressions calculated before and after treatment so that the results of the whole analysis may be open to some question. As indicated by SNEDECOR [1946], this bias is not serious unless the error in X is large as compared to the range of the data.

As a final point, interpretation of the results of the mathematical methods to be presented will, like any similar methods, require the use of some judgment and experience. In addition to the statistical estimates, for example, the investigator should take into consideration the range of recorded data. If the period of record has been unusually uniform in climatic variations, as may happen in small samples, it may be desirable to continue the calibration period until a more reasonable range of data is available, as through the occurrence of wetter or drier years.

For most purposes, it is considered advisable to have the periods before and after treatment as nearly the same in length as possible, so as to give both an equal opportunity to be exposed to variations in climate and other factors. The method to be outlined is planned to meet this objective as nearly as possible. After estimating the calibration period with its help, however, the investigator should keep in mind that it is better to make this set of records a little longer than necessary, rather than much too short. Obviously, additional calibration data cannot be obtained after the treatment has once been applied. But with adequate pre-treatment records on hand, the study can be terminated after treatment whenever the effects have been demonstrated to a satisfactory degree.

The analytical method--In order to develop a background for the statistical method which is to be employed, let us deal first with an experiment employing two watersheds, in which the whole study has been completed and the data analyzed. As outlined in the author's earlier article on this subject, the method of covariance analysis is used to test the significance of the differences between averaged results for two periods of time (before and after treatment of one watershed), after the data have been adjusted by regression to minimize the influence of variations in climate and other extraneous factors.

When both periods contain the same number of observations, and when a single regression coefficient is common to both regression equations in this kind of analysis, the variance of the difference between these adjusted means may be estimated by the equation

$$s_d^2 = s_{y\cdot x}^2 \left[2/k + (\bar{x}_1 - \bar{x}_2)^2/Sx^2 \right] \ldots\ldots\ldots\ldots\ldots\ldots (1)$$

where s_d^2 is the desired variance of the difference between adjusted means; $s_{y\cdot x}^2$ is the variance of estimate (squared standard error of estimate [EZEKIEL, 1941]; k is the number of observations in each set of data (as years of record, when annual water yields are being considered); $\bar{x}_1$ is the mean of the data for the control watershed (independent variable, X) for the period before treatment; $\bar{x}_2$ is the corresponding mean after treatment; and Sx^2 is the summed squared deviations of the independent variable (within periods). (Symbols and terminology are those of SNEDECOR [1946].) Using data from the author's earlier article [WILM, 1944], (see Tables 1 and 2),this equation may be set up in numerical terms as

$$s_d^2 = 3954.4 \left[2/5 + (904.0 - 602.0)^2/498{,}800 \right] = 2304.8$$

Then the standard error of the adjusted difference is

$$s_d = \pm \sqrt{2304.8} = \pm 48.0 \text{ ac ft of water;}$$

and

$$t = d_y/s_d = (\hat{Y}_1 - \hat{Y}_2)/s_d = (719.7 - 492.3)/48.0 = 4.74$$

In the last equation, $\hat{Y}_1$ and $\hat{Y}_2$ are the adjusted estimates of water yields from the treated watershed. When the above t is squared, the result should equal the value of F calculated in the covariance analysis

$$F \doteq 88{,}708.3/3{,}954.4 = 22.43 = (4.74)^2 \text{ (approximately)}$$

It follows that the square of the smallest significant difference may be estimated by multiplying (1) by F on any desired odds.

$$d^2 = s_{y\cdot x}^2 F \left[2/k + (\bar{x}_1 - \bar{x}_2)^2/Sx^2 \right] \ldots\ldots\ldots\ldots\ldots\ldots (2)$$

Table 1--Annual discharges from snow, Watersheds A and B

Year	Before treatment: Discharge: Watershed B X	Before treatment: Discharge: Watershed A Y	Year	After treatment: Discharge: Watershed B X	After treatment: Discharge: Watershed A Y
	ac ft	ac ft		ac ft	ac ft
1940	950	540	1946	580	690
1941	1060	650	1947	250	360
1942	540	380	1948	710	700
1943	1200	780	1949	920	850
1944	780	590	1950	550	520
Sum (SX, SY)	4520	2940	Sum (SX, SY)	3010	3120
Mean ($\bar{x}_1$, $\bar{y}_1$)	904.0	588.0	Mean ($\bar{x}_2$, $\bar{y}_2$)	602.0	624.0
			Grand sum (for 10 yrs)	7530	6060

Table 2--Analysis of covariance; effect of timber-cutting on water yield from Watershed A, adjusted to mean yield from Watershed B

Line	Source	Degrees of freedom	Corrected squares and products: Sx^2	Corrected squares and products: Sxy	Corrected squares and products: Sy^2	b[a]	Degrees of freedom	Errors of estimate: Sum of squares	Errors of estimate: Mean square
1	Total	9	726,810	288,920	231,240		8	116,389.1[b]	14,548.6
2	Periods	1	228,010	-27,180	3,240		1	88,708.3	88,708.3[c]
3	Error, (1) - (2) (within periods)	8	498,800	316,100	228,000	0.6337	7	27,680.8[b]	3,954.4
4	Period 1	4	258,920	139,140	86,280	0.5374	3	11,508.1[b]	. . . . .
5	Period 2	4	239,880	176,960	141,720	0.7377	3	11,176.2[b]	. . . . .
6	Sum, (4) + (5)	. .	498,800	316,100	228,000	. . . .	6	22,684.3	3,780.7
7	Within periods, (3)	. .	498,800	316,100	228,000	. . . .	7	27,680.8	. . . . .
8	Diff. (7) - (6)	. .	. . . . .	. . . . .	. . . . .	. . . .	1	4,996.5	4,996.5[d]

[a] $b = [Sxy/Sx^2]$ = regression coefficient. [b]Sum of squares = $\{Sy^2 - [(Sxy)^2/Sx^2]\}$. [c]Treatment significant; odds less than 0.01 that the difference between periods is due to chance alone. [d]Not significant.

Examination of (1) or (2) shows a quantity which is needed in the particular problem to be discussed in this paper: k, the number of observations in each of the two periods before and after treatment. As employed up to this point, k is a known quantity; for our new purpose it is the unknown which is to be estimated.

Before we can make use of (2) for this purpose, it will be necessary to provide a reasonable estimate for the probable difference between the means of X before and after treatment.

$$(\bar{x}_1 - \bar{x}_2)^2 \longrightarrow 2 s_x^2 F/k$$

where s_x^2/k is the variance of a sample mean of X, and F may be chosen at any desired level of significance.

After this term is inserted and Sx^2 is replaced by $2\ (k - 1) s_x^2$, (2) can be simplified and solved, in part, for k

$$k = (s_{y.x}^2 F/d^2)\left\{2 + F/(k - 1)\right\} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (3)$$

If desired, (3) may be treated as a quadratic equation and solved entirely for k.

$$k = s_{y \cdot x}^2 F/d^2 + \left[1 \pm \sqrt{(4 s_{y \cdot x}^2 F/d^2)(s_{y \cdot x}^2 F/d^2 - 1 + F) + 1}\right]/2 \quad \ldots\ldots\ldots \text{(3a)}$$

In working out either (3) or (3a), the magnitude of F has to be based on a rough estimate of the number of degrees of freedom that will be available for the within-periods term in the analysis of final results, using the formula

$$\text{Degrees of freedom} = 2(k - 1) - 1$$

Thus even (3a) has to be solved by one or more approximations. For this reason it is suggested that a sufficiently reliable estimate of k may be obtained most easily by trial solutions of (3), with the assistance of values given in Table 3 for the factor F (2 + F/k - 1), corresponding to various magnitudes of k.

Table 3--Values of F and of F(2 + F/k - 1), for use in the solution of (3)

k	F[a]	$F\left[2 + F/(k-1)\right]$
3	10.13	71.52
4	6.61	27.76
5	5.59	19.01
5	5.12	15.46
7	4.84	13.60
8	4.67	12.47
9	4.54	11.67
10	4.45	11.08
11	4.38	10.69
12	4.32	10.32
13	4.28	10.10
14	4.24	9.88
15	4.21	9.68
20	4.10	9.10

[a] For this table, F is taken at the 0.05 level, on 2(k - 1) - 1 degrees of freedom. Where k is 5, for example, the tabular value of F is associated with 2(5 - 1) - 1, or 7 degrees of freedom.

As we look at (3) and (3a) again, it will be noticed that the magnitude of d^2 cannot be calculated from calibration-period data, but must be chosen by the investigator on a logical basis. Because this quantity plays an important part in the equations, it may be worth a little further discussion. In any study it may be possible, by obtaining a sufficient number of observations, to demonstrate that even a very small, though real difference due to treatment is greater than might have occurred by chance. But there may be a point below which such an effect would have little practical significance; and it would involve unnecessary and perhaps excessive costs to design and carry through an experiment that would be elaborate and efficient enough to detect it. Hence it is necessary to decide what constitutes a treatment effect large enough to warrant practical consideration. This question may become complex, particularly when the cost of treatment must be charged against its benefits. For our purpose, however, it can only be remarked that the investigator must establish some quantitative value for this smallest worth-while difference due to treatment. And then, of course, it is also necessary to decide on the odds which, according to our desires, are required to show that this minimum difference is greater than might have occurred purely as a result of chance variations.

Examples of the method--To show how this method may be employed in watershed investigations, it will be desirable to describe several numerical cases. In the first one we shall use the data shown for the pair of watersheds in Tables 1 and 2, assuming that we have completed only the period before treatment and wish to estimate, after this five-year period, whether the watersheds have been calibrated long enough.

In setting up the numerical data for (3), the minimum worth-while difference d is taken at 20 per cent of the average yield from Watershed A before treatment: $d = 588.0 \times 0.20 = 117.6$, or about 120 ac ft. Our estimate of $s_{y \cdot x}^2$ is taken from Errors of estimate, in line 4 of Table 2: $s_{y \cdot x}^2 = 11{,}508.1/3 = 3836.0$. With these data available we can solve (3) by means of two or three steps. First is computed $s_{y \cdot x}^2/d^2 = 3{,}836.0/14{,}400 = 0.266$. If, as a preliminary guess, we set a value of seven years for k, we can enter Table 3 at this line and read the quantity 13.60 in the right-hand column. Then our first trial solution of 3 is $k = (0.266)(13.60) = 3.62$ years. Because this is considerably lower than our preliminary guess of seven years, we may make a second solution, taking an intermediate estimate of five years for k.

$$k = (0.266)(19.01) = 5.06 \text{ years}$$

As this agrees very well with our revised estimate, we can say that about five years should be required in each of the periods, before and after treatment, in order to demonstrate that a treatment effect of 120 ac ft is greater than might be expected by chance.

Now, using all the data available after conclusion of the study, it may be interesting to see whether our estimate was reasonably correct. Returning to (1), we have already calculated a standard error of the difference between adjusted means

$$s_d = \pm \sqrt{2304.8} = \pm \text{ 48.0 ac ft of water}$$

based on the data obtained for the whole study. If we should require odds of 19:1 in testing the significance of the differences between adjusted means, the least significant difference would be estimated as

$$d'_y = 48.0 \times t_{0.05} = 48.0 \times 2.365 = 113.5 \text{ ac ft}$$

on seven degrees of freedom. As this quantity is smaller than the 120 ac ft which was set as the smallest worth-while difference, our estimate of five years per period was satisfactory.

After the treatment has been applied and a part of the subsequent records are on hand, it should be desirable to make one or more preliminary tests of the results. The purpose of these tests would be to make sure that the post-treatment period of record is not prolonged unnecessarily. They may be performed, of course, simply by carrying through covariance analyses similar to that shown in Table 2.

For our second example we can use the data from a watershed investigation being conducted by the Rocky Mountain Forest and Range Experiment Station, for which records for a number of years are available. At this stage in the study we want to decide whether enough calibration data have been obtained, and approximately when treatments may be applied. Of the two watersheds to be considered, the one to be treated (Y) is 1.11 sq mi, while one of two which may serve as climatic controls is much larger, 32.8 sq mi (Watershed X). Both watersheds are located in the lodgepole pine and spruce-fir forest types near the Continental Divide in Colorado. As the smaller area is contained within the larger one, the yields of Watershed X were adjusted to remove those of Watershed Y when the data were prepared for analysis.

In Table 4 and Figure 1 are presented the annual water yields (expressed as depth in inches over the watershed) for both of these areas, covering a period of seven water years which ended on September 30, 1946. These particular years since Watershed Y was placed under control seem to form a representative sample of a longer period, as judged by comparing the 1940-1946 data for Watershed X with data from the same area for the 12-year period, 1935-1946. The seven-year average yield was about 97 per cent of that for the longer period, and both exhibited comparable variations in yield. Thus it may be considered that our seven-year record for the two watersheds together contains a desirable set of data on water yields. This conclusion is supported by Figure 1, which indicates that the relation between yields of the two watersheds may be expressed in a satisfactory manner by a straight line.

Table 4--Annual water yields, Watersheds X and Y

Water year	Yield--Inches depth	
	Watershed X	Watershed Y
	in	in
1940	10.88	7.21
1941	13.72	11.85
1942	15.75	14.86
1943	17.38	14.45
1944	14.31	9.50
1945	15.43	11.47
1946	14.87	12.79
Total	102.34	82.13
Average	14.620	11.733

From the values in Table 4 was calculated the regression of yields from Watershed Y on those from Watershed X (see Table 5), together with the necessary errors of estimate. As indicated by the correlation coefficient (+0.872), these yields are correlated well enough so that we need not be concerned over the magnitude of any bias that may be associated with errors in both Y and X.

In proceeding toward the solution of (3) in this study, the minimum worth-while difference is set at an annual yield of 2.0 inches, only 17 per cent of the average pretreatment water yield of watershed Y. Using this figure and our calculated value for the variance of estimate, we can obtain

$$s_{y \cdot x}^2/d^2 = 2.0917/4.0 = 0.5229$$

from which, by one or two approximate solutions of (3), we arrive at about seven years as our best estimate for the minimum duration of the periods before and after treatment.

In our last example we shall use a problem that is encountered during the organization of new research. Assuming that no watershed study has yet been started, we want to estimate whether a

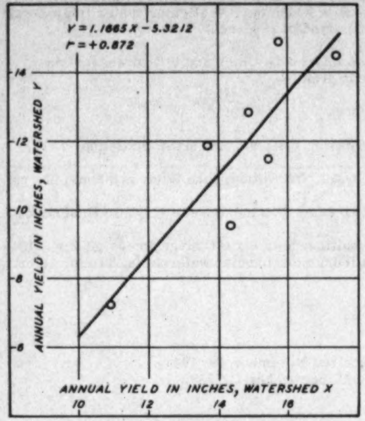


Fig. 1--Regression of yield from Watershed Y on yield from Watershed X

given pair of drainage units, tentatively chosen for experimental control, is likely to produce results of practical significance in a reasonable period of time. We have no measured data for the areas in question, but have some knowledge as to the local magnitude of variables to be studied, based on stream-flow records that have been taken in the same general locality. We can state within reasonable limits, for example, that the mean annual yield of areas like these is roughly seven inches of water; and that, within a short period of time such as a decade, yields in individual years are likely to vary from about three to 12 inches. Judging from past experience in similar watershed investigations, too, we can say that the yields of contiguous and relatively small areas are likely to be highly correlated, and have decided to set a probable value of +0.90 for their correlation coefficient.

From our knowledge of the probable range in yields we can make an estimate of their standard deviation [SNEDECOR, 1946, pp. 97-98]. Assuming that this range is likely to occur within a ten-year period.

$$s = 12 - 3/3.08 = 2.92$$

Table 5--Analysis of water yields, Watersheds X and Y

Degrees of freedom	Corrected squares and products			Degrees of freedom	Errors of estimate	
	Sx^2	Sxy	Sy^2		Sum of squares	Mean square
6	24.5068	28.5866	43.8041	5	10.4585	2.0917

Regression coefficient (b) = $Sxy/Sx^2 = 28.5866/24.5068 = +1.1665$

Regression equation = $\hat{Y} = \bar{y} + b\,(X - \bar{x})$

$\hat{Y} = 11.733 + 1.1665\,(X - 14.620)$

$= 1.1665X - 5.3212$

Correlation coefficient (r) = $\sqrt{(Sxy)^2/(Sx^2)(Sy^2)} = +\ 0.872$

from which the estimated variance s^2 is calculated to be 8.54.

Finally, we can obtain an approximate value for the variance of estimate ($s_{y\cdot x}^2$) by using our knowledge as to the probable degree of correlation between yields of contiguous watersheds. The magnitude of $s_{y\cdot x}^2$ may be estimated by the equation

$$s_{y\cdot x}^2 = (1 - r^2)\, s^2\left\{2\,(k - 1)/\left[2(k - 1) - 1\right]\right\} \ldots\ldots\ldots\ldots\ldots\ldots \quad (4)$$

Because k is the unknown we want to estimate, we shall have to employ a rough first approximation for its value; or, if it is expected to be fairly large, the portion of (4) contining k may be neglected. If, however, we estimate k as about 5 and r^2 as about 0.81,

$$s_{y\cdot x}^2 = (1.00 - 0.81)(8.54)8/7 = 1.85$$

Our only remaining requirement is an estimate of the minimum worth-while difference. By setting this at about 20 per cent of the mean annual yield, or about 1.4 inches of water, we can calculate

$$s_{y\cdot x}^2/d^2 = 1.85/1.96 \quad 0.95$$

Using this figure and the data in Table 3, our best estimate for the minimum period is about ten years.

In its application to new studies like this, of course the method has decided limitations; its usefulness depends very largely upon the precision with which the investigator can estimate local conditions, and upon his breadth of experience in this kind of research.

Acknowledgments--Acknowledgment is made to Andrew G. Clark and to Laurence Allen for their suggestions on mathematical aspects of this article.

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(Communicated manuscript received November 18, 1947;
open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

SOME INFLUENCES OF FROST PENETRATION ON THE HYDROLOGY OF SMALL WATERSHEDS

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Abstract--An example of the influence of frost penetration on hydrologic as well as agronomic problems is presented and discussed. The data indicate frost penetration retards percolation thereby retaining water in the soil profile that normally would drain. Frost penetration through its influence on percolation and resultant available storage may affect infiltration for a considerable time after the frost period.

Introduction--In the course of an investigation on soil moisture and frost penetration a rather unique combination of circumstances was observed. Early in 1947, the frost penetration appeared to influence the soil-water relationships during and for a considerable time after the period of frost. The peculiar circumstances prior to and during this period are briefly described. It is believed this report will be of interest to hydrologists and particularly those engaged in agricultural hydrology and in studies bearing on soil and water conservation. It presents an example of how closely frost penetration can influence hydrologic as well as agronomic problems.

Observations were made on the United States Soil Conservation Service Hydrologic Experiment Station, Coshocton, Ohio. This station is maintained in cooperation with the Ohio Agricultural Experiment Station, Wooster, Ohio.

Review of past work--After conducting hydrologic studies on small watersheds for over a decade, it has been observed that along with the variations in water disposal from year to year, there is also a certain amount of seasonal regularity in the depletion of soil water by various means. In this region percolation to the water table occurs mostly during late winter and spring. Evapotranspiration rates are highest during late spring and summer. Runoff peaks from small watersheds are highest from May to September [DREIBELBIS, 1947; DREIBELBIS and HARROLD, 1945]. Soil-moisture measurements also have shown that the amount of water stored in the soil profile usually reaches a maximum in February, March, or April. The exact time depends largely on weather conditions. Soil water usually reaches a minimum in August, September, or October, again depending largely on the weather. From October through March the trend of soil-moisture storage is upward. From April through September it is downward. Hydrologic data confirming these trends were previously presented in several papers by the author listed in the References.

Studies of frost penetration by various investigators have dealt largely with the influence of land use and vegetative cover on depth of frost penetration and upon the effects of frost depth on surface runoff. Frost-penetration records have been obtained at this station for eight years, 1940-1947, yet only once, 1940, did frost penetration affect runoff. The rest of the time the ground was not frozen when surface runoff occurred. If there was frost, the precipitation was generally in the form of snowfall. Consequently, there was no surface runoff.

Experimental procedure--The depth of frost penetration was determined by finding the line of demarcation between unfrozen soil and the ice crystals in frozen soil. A mattock or a large diameter twist drill was used for this purpose. Details of the method were previously presented by POST and DREIBELBIS [1942]. Percolation was measured by monolith lysimeters 0.002 acre in area and with a profile depth of eight feet. These lysimeters have been fully described by KOHNKE, DREIBELBIS, and DAVIDSON [1940]. Precipitation was measured by recording rain gages. Daily soil-moisture curves were worked out from data obtained from the weighing lysimeters and checked with the gravimetric method. Details of this procedure were previously described by HARROLD and DREIBELBIS [1945].

Results--January 1947 was characterized by abnormally high temperatures and high precipitation mostly in the form of rain. The mean temperature for January 1947 was 34.3°F and the normal for a 23-year period in Coshocton, Ohio is 29.3°F [FISHER, 1941]. The normal and monthly precipitation values for 1947 appear in Figure 1. Due to the high precipitation and to temperatures above the freezing point the soil moisture storage was increased to such an extent that the profile was practically saturated by the end of the month. This situation was confirmed by sustained high

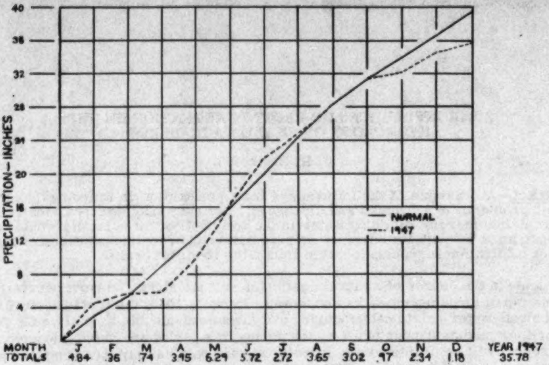

Fig. 1--Monthly precipitation, 1947 and normal

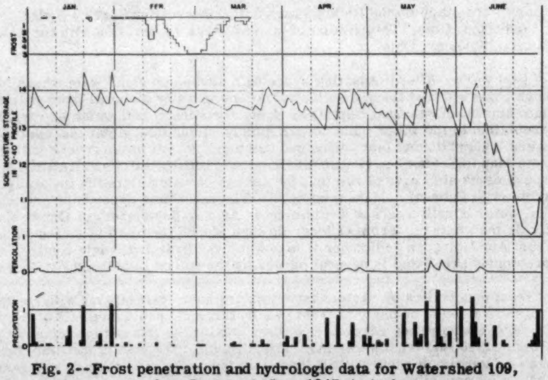

Fig. 2--Frost penetration and hydrologic data for Watershed 109, meadow, January to June 1947, in inches

percolation rates in monolith lysimeters of a profile depth of eight feet (see Fig. 2). Data on frost penetration, percolation, precipitation, and soil-water storage are given for the Muskingum silt loam on Watershed 109, from January to June 1947 in Figure 2.

February was characterized by abnormally low precipitation, 0.36 inches, and temperatures averaging eight degrees below normal. The latter condition resulted in frost penetration of the soil to a depth of five inches on meadow watersheds (see Fig. 2). In February and March 1947, as long as the deep frost prevailed, the soil profile was practically saturated. Percolation almost stopped. It reached a minimum rate of 0.006 inch per day. As soon as the frost left the soil, air penetrated the soil and more rapid percolation, 0.012 inch per day, ensued. Eight days later, March 30, this rate increased to 0.069 inch per day although only 0.34 inch precipitation fell during this period. The high soil moisture during the period of frost was thereby confirmed. Thus, in spite of abnormally low precipitation, both in February, 0.36 inch, and March, 0.74 inch, the soil profile was practically saturated during this period.

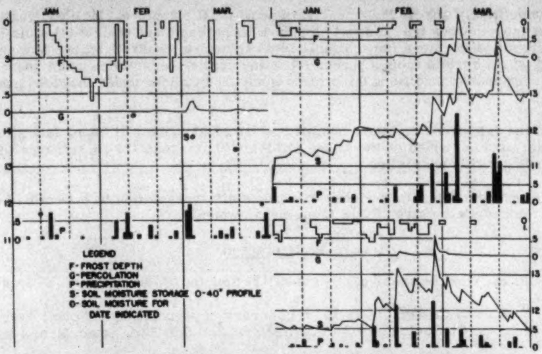

Fig. 3--Frost penetration and hydrologiv data for Watershed 109, January to March 1940, wheat; 1945, meadow; 1946, wheat; in inches

In April, May, and early June, the precipitation was above normal. A high soil-moisture content continued throughout this period. The soil-moisture content was so high that oats and corn planting were considerably delayed in many locations in Ohio. In some areas oats was not planted at all owing to excessive soil moisture for tillage operations. On many farms, corn planting was delayed to such an extent that the crop was not mature when the first frost appeared and corn yields were considerably below normal.

The influence of frost penetration in other years is shown in Figure 3. The major factors influencing percolation are soil-moisture content, precipitation, and frost penetration when soil type and land use are constant. When soil moisture is sufficiently high for water to percolate through the soil, frost can have a great influence on percolation. This situation appeared in the years 1940, 1945, 1946, and 1947 (see Fig. 2 and 3). During the other years 1941, 1942, 1943, 1944, the soil-moisture content was not sufficiently high in the winter to start percolation and frost had little influence on percolation until in March. The old saying "Nature abhors a vacuum" can be applied to the influence of frost penetration on percolation. Frozen soil admits so little air that percolation can proceed only very slowly because a partial vacuum would be produced if it were rapid. Here the type of frost structure is also very influential. The concrete type of frost POST and DREIBELBIS, 1942 is very dense. Little air penetrates the soil and percolation practically ceases. In the more porous type of frost (the honeycomb and stalactite structures) more air may enter than in the concrete type and percolation is likely to be somewhat greater. The honeycomb and stalactite structures generally occur in shallow frost and principally in meadow and pasture covers. The concrete type is associated with deep frost and generally in fallow or cultivated land [POST and DREIBELBIS, 1942]. Thus, meadow and pasture covers and shallow frost conditions will tend to retard percolation less than under corn, wheat or fallow conditions where the frost is deep.

The influence of frost penetration upon infiltration has been reported by numerous investigators largely from the viewpoint of the effect of frost upon surface runoff. In the study reported by SCHIFF and DREIBELBIS [1949] the influence of the soil moisture on infiltration was illustrated.

The hydrologic data and frost reported in this paper indicate the influence of frost upon infiltration through retention of the water in the profile. The available storage capacity of the soil is thereby kept at a minimum. Consequently, infiltration rates are lower. In 1947 this condition proved detrimental to crops because of excessively high soil-moisture conditions for tillage operations. In some years precipitation after the frost period might be very low. Then a full reservoir of soil water early in the growing season would sustain and promote better crop growth than otherwise.

Land use definitely affects frost penetration and frost structure. Meadow, pasture, and woodland favor the more porous type of frost structure (honeycomb and stalactite) while cultivated land favors deep frost and the more impermeable frost structure (concrete type). More available soil-water storage space results from a porous frost and greater percolation under these conditions. Thus, under comparable conditions infiltration would be enhanced under meadow, pasture, and woodland covers.

Although the data presented were obtained on the Muskingum silt loam, data similar to those in Figure 2 were also obtained on the Keene silt loam (Watershed 123), a soil type with distinctly different hydrologic characteristics.

Acknowledgment--The author wishes to express his appreciation to Robert E. Youker, Agricultural Aid, for assistance given in the frost measurements.

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(Communicated manuscript received April 29, 1948;
open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

RECOVERY OF GROUND-WATER SUPPLIES BY PUMPING FROM WATER-TABLE PONDS

Henry N. Halberg and Claude M. Roberts

(Published by permission of the Director, Geological Survey, United States Department of the Interior, and the Commissioner of Public Works, Commonwealth of Massachusetts.)

Abstract--This paper summarizes a study made to determine whether ground-water storage in the glacial deposits in the vicinity of Fresh Pond, a water-table pond used as part of the public supply of the City of Cambridge, Massachusetts, is available to augment the supply from surface-water reservoirs.

Test wells were drilled; water level, well and test boring, and quality of water data, and data on inflow and outflow from Fresh Pond were collected, and seismic profiles were run, to determine whether the glacial deposits in the vicinity of the pond contribute appreciable quantities of water to the pond. The data show that water levels in certain wells fluctuate with the level of Fresh Pond and that certain other water levels do not, indicating ground-water contributions to the pond when it is at low stages and increases in ground-water storage when pond stages are high. The largest changes in ground-water storage occur in the glacial deposits underlying the banks to the west and northwest of the pond. Mineral characteristics of the well and pond waters offer no evidence as to the direction of ground-water flow.

Introduction

Investigations of the ground-water resources of Massachusetts were begun by the Geological Survey in 1938 in cooperation with the Massachusetts Department of Public Works. The work was started in the northern part of Middlesex County in the area around Lowell and in the Aberjona Valley. Two papers by BRASHEARS [1941 and 1944], concerning the work in this area were presented before the New England Water Works Association. Included in this work was a program of water-level measurements, the results of which have been published in UNITED STATES GEOLOGICAL SURVEY [1939-1944] Water-Supply Papers. This work was done as the beginning of a state-wide investigation of ground-water resources.

In 1944 studies were begun of the ground-water resources of southeastern Middlesex County, particularly around Fresh Pond in Cambridge, the field work for which was done in 1944 and 1945. These studies were made at the request of the Massachusetts Department of Health for the purpose of determining whether the City of Cambridge could meet its legal requirements for water supply. According to State law, the cities and towns in the Metropolitan Water District not taking water from the metropolitan supply must be able to maintain a supply of not less than ten per cent more than the average consumption for the preceding three years. The United States Geological Survey was requested to make a study to determine the potential supply of ground water available in the glacial sands and gravels surrounding Fresh Pond.

As part of this study all available records of wells, test borings, and other subsurface explorations in the area were collected. Test wells were driven to expand these data. Geologic investigations carried out by the Geologic Branch of the United States Geological Survey consisted of preparation of a geologic map of the area, supplemented by seismic investigations made by Daniel Linehan to determine the thickness of the overburden.

Geography

Topography and drainage of the area--The City of Cambridge is near the center of the Boston metropolitan area, and had a population of 111,124 in 1945. The city covers an area of about seven square miles, with an average elevation of about 22 ft above sea level. This report covers an area of slightly less than six square miles, situated in the western part of Cambridge and adjacent parts of Arlington, Belmont, and Watertown (see Fig. 1). The area around Fresh pond, where this study

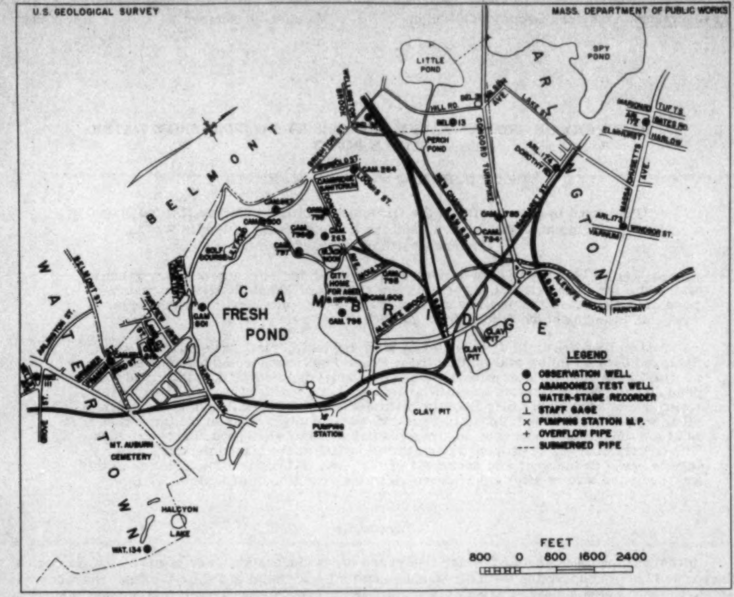
U.S. GEOLOGICAL SURVEY
MASS. DEPARTMENT OF PUBLIC WORKS
LITTLE POND
SPY POND
BELMONT
ARLINGTON
CAMBRIDGE
WATERTOWN
FRESH POND
PERCH POND
CLAY PIT
MT. AUBURN CEMETERY
HALCYON LAKE
PUMPING STATION
CONCORD TURNPIKE
LAKE ST.
MASSACHUSETTS AVE.
ALEWIFE BROOK PARKWAY
BELMONT ST.
ARLINGTON ST.
WAT.134
LEGEND
● OBSERVATION WELL
○ ABANDONED TEST WELL
WATER-STAGE RECORDER
⊥ STAFF GAGE
× PUMPING STATION M.P.
+ OVERFLOW PIPE
--- SUBMERGED PIPE
FEET
800 0 800 1600 2400

Fig. 1--Map of Fresh Pond and vicinity showing locations of observation wells, abandoned test wells, and pond water-level measuring points

was mainly concentrated, is part of the Mystic River drainage system. The pond is on the upper reaches of Alewife Brook, a tributary of the Mystic River. Much of the area in the immediate vicinity of Fresh Pond is quite flat, being only about ten to 20 ft above sea level, and most ot the remainder is less than 50 ft above sea level.

Climate--The climate of Cambridge is typical of that of Metropolitan Boston. Climatological records have been collected at nearby Boston since 1818, and the data used in this report are based on records of the United States Weather Bureau station of that city.

The normal temperature at Boston is 49.6°F., based on records for the period 1875 to 1921. The lowest annual mean temperature was 46.6°F., in 1875; the highest, 53.0°F., in 1931. The lowest monthly normal, 27.9°F., occurs in January and the highest, 71.7°F., in July. The range in temperature is 122°F.; the lowest, -18°F., occured in February 1934, the highest, 104°F., in July 1911.

The normal annual precipitation at Boston is 40.14 inches, based on a 50-year period of record, 1878-1927. During the period 1818-1946 the lowest yearly amount, 27.20 inches, occurred in 1822; the highest, 67.72 inches, in 1863. The monthly precipitation is fairly evenly distributed during the year; the lowest monthly normal precipitation, 2.89 inches, occurs in June and the highest, 3.62 inches, in August and January, respectively. The driest month on record was October 1924, with 0.06 inches, and the wettest, July 1863, with 12.38 inches.

Geology

The Fresh Pond area lies within the Boston basin, a lowland characterized by low hills of glacial drift and flat plains of glacial outwash and bounded on the north and west by a pronounced escarpment. The Boston Basin is presumably a down-faulted block of old rocks which are covered

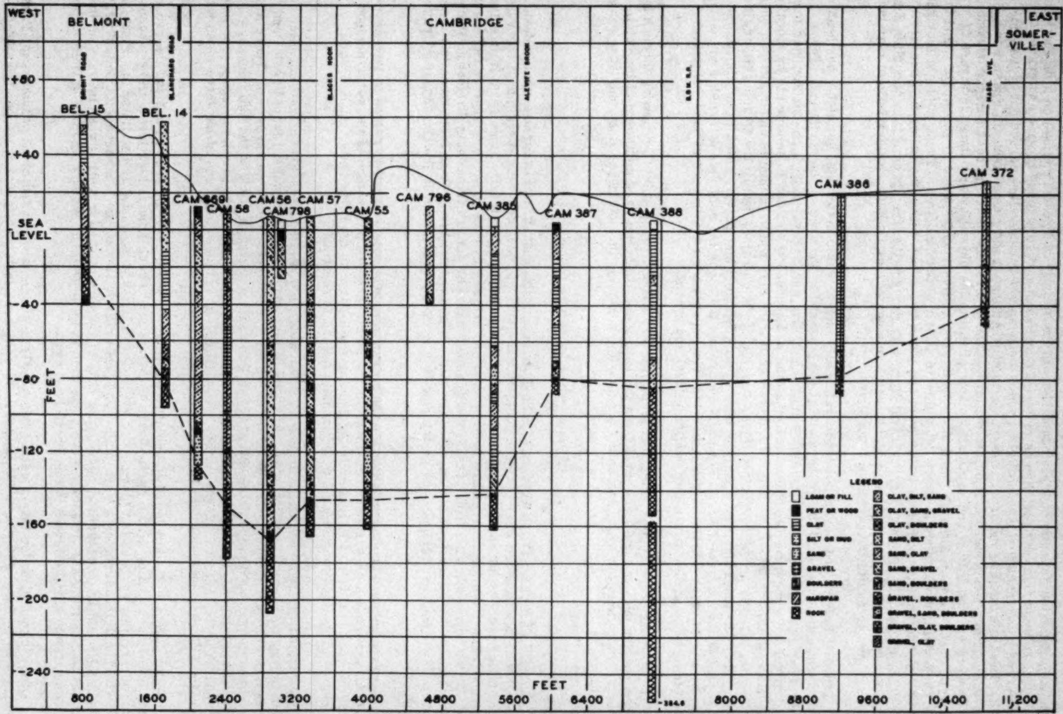
WEST
BELMONT
CAMBRIDGE
EAST
SOMER-VILLE
BRIGHT ROAD
BLANCHARD ROAD
BLACKS NOOK
ALEWIFE BROOK
B.& M. R.R.
MASS. AVE.
BEL. 15
BEL. 14
CAM 369
CAM 58
CAM 56
CAM 798
CAM 57
CAM 55
CAM 796
CAM 385
CAM 387
CAM 388
CAM 386
CAM 372
+80
+40
SEA LEVEL
-40
-80
-120
-160
-200
-240
FEET
800
1600
2400
3200
4000
4800
5600
6400
8000
8800
9600
10,400
11,200
FEET
-254.6
LEGEND
LOAM OR FILL
PEAT OR WOOD
CLAY
SILT OR MUD
SAND
GRAVEL
BOULDERS
HARDPAN
ROCK
CLAY, SILT, SAND
CLAY, SAND, GRAVEL
CLAY, BOULDERS
SAND, SILT
SAND, CLAY
SAND, GRAVEL
SAND, BOULDERS
GRAVEL, BOULDERS
GRAVEL, SAND, BOULDERS
GRAVEL, CLAY, BOULDERS
GRAVEL, CLAY

Fig. 2--Geologic Section north of Fresh Pond, Cambridge, Massachusetts, from Belmont to Somerville, along line of Metropolitan District Water Supply Commission borings

with unconsolidated deposits of Pleistocene age. In the vicinity of Fresh Pond, which is in the western part of Cambridge, the Pleistocene material consists of irregular beds of sand and gravel interspersed with lenses and layers of clay, which lie in an old buried valley carved out of the bedrock. The valley enters the area from the north near Arlington and passes beneath Fresh Pond, where its bottom lies about 160 ft below sea level. The general shape of the rock surface beneath Fresh Pond and the type of unconsolidated material filling the old valley are shown in Figure 2. It can be observed that, although some of the beds are relatively widespread in extent, they tend to interfinger with one another or pinch out within relatively short distances. These geologic features tend to impede the flow of ground water from one place to another within the valley fill, but fortunately, by forcing the ground water to travel along circuitous paths, they permit it to percolate downward and recharge the lowermost aquifers, which in many places lie beneath thick beds of clay but in others are accessible to recharge from above.

Much of Fresh Pond is surrounded by relatively impermeable materials, but highly permeable glacial outwash lies around the western end of the pond and in the area to the northwest. These water-bearing materials are underlain by ground moraine and beds of clay which extend under Fresh Pond to form an impermeable bottom. To the east and south, Fresh Pond is hemmed in by morainal till. Beneath this material lies a thick impervious marine clay which in turn is underlain by a sheet of till lying directly on the bedrock surface. In front of the moraine, toward the Charles River, the shallow deposits are composed chiefly of beds of outwash sand and gravel, which lie at elevations above that of the surface of Fresh Pond. The morainal till along the east side of Fresh Pond is overlain by a relatively thick deposit of impervious clay interbedded with layers of sand and gravel, which in turn is mantled with a thin veneer of sand and gravel.

Thus, much of the ground water that seeps into Fresh Pond reaches it after flowing through the interconnected beds of sand and gravel from the catchment areas west and northwest of the pond. However, other smaller sandy catchment areas to the north and the southwest also contribute some recharge.

Ground-water hydrology

Sources of water--Fresh Pond is a kettle-hole reservoir, that is, it was formed when a block of stagnant ice, embedded in an outwash plain, was melted. No surface water flows into it from the surrounding area; under natural conditions all the water that reaches it falls directly on it or enters it as underflow from the surrounding beds of sand and gravel when the reservoir stage is relatively low. Water reaches it also from outside the drainage basin by conduit from Stony Brook and Hobbs Brook Reservoirs in the town of Lincoln. The water which enters Fresh Pond from these two reservoirs passes over an overflow weir on the pipe line from the reservoirs to the Cambridge filtration plant.

Water is pumped from Fresh Pond to the filtration plant, from which it is raised to Payson Park Reservoir, a high-pressure reservoir, in Belmont. A small percentage of the water pumped from the pond is returned to it after being used as wash water in the filter beds. Also, a small amount of water used in washing the sedimentation beds is returned.

Intermittently from March 23 to June 14, 1945, water was wasted from Fresh Pond into Alewife Brook through a 24-inch overflow pipe. At times of high stage, water is wasted from Fresh Pond through two 12-inch pipes into Lily Pond, from which it discharges into a storm sewer. There are also three 12-inch waste pipes connecting Fresh Pond with a drainage ditch on the golf course, northwest of the pond, from which water was originally wasted to Alewife Brook through a pipe line and culvert under Concord Avenue, which is now filled. At various times one of the three pipes was opened and discharged water into Lily Pond through a drainage ditch.

Water levels --To determine the relation between the water level in Fresh Pond and the ground water in the surrounding areas, particularly to the north of Fresh Pond, water levels were measured in 20 observation wells and at four ponds; Fresh Pond, Black's Nook, Little Pond, and Spy Pond (see Fig. 1). These measurements were made in 12 test wells drilled by the Massachusetts Department of Public Health, in three wells drilled by the City of Cambridge, and in wells owned by various industries and individuals. Readings were made also at three wells in Woburn and one in Wilmington, for comparison with the levels in the Fresh Pond area.

Observations of water levels were made weekly in all these wells and at all ponds except Fresh Pond, and wells Cam. 667, Cam. 798, and Wil. 44. On these latter wells and at the pond, water-stage recorders were installed to furnish continuous records of stage. Water-stage recorders were installed on wells Cam. 667 and 798, as these two wells end in the two major aquifers in the area. Well

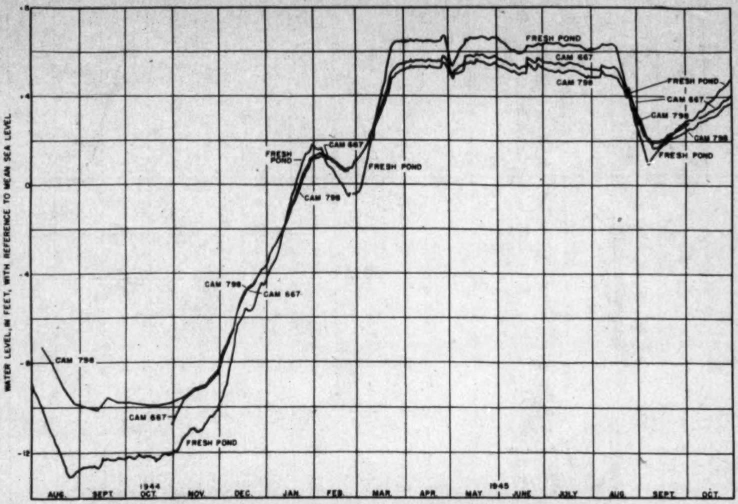

Fig. 3--Hydrographs based on daily measurements, showing the fluctuations of water levels in Fresh Pond and in wells Cam. 667 and Cam. 798

Cam. 798, 31 ft deep, penetrates the shallow gravel beds and well Cam. 667, 127 ft deep, ends in the deeper gravels that lie beneath layers and lenses of clay. Examination of the hydrographs of these two wells and that of Fresh Pond (see Fig. 3) indicates that fluctuations in the water level in both the wells closely follow those in Fresh Pond.

The trend in water levels in wells Arl. 172, Arl. 174, Cam. 667, and Cam. 798 is shown on another hydrograph (see Fig. 4). Well Arl. 174, 176 ft deep, is slightly more than a mile north of the pond. Well Arl. 172 is a shallow well 36 ft deep, located about 7,500 ft north of the pond, on Massachusetts Avenue. The water levels in three of these wells follow Fresh Pond fluctuations, but at Arl. 172 the pattern is different.

The test and observation wells were located in lines extending radially from Fresh Pond. Profiles of water levels along these lines when the pond level was being raised or lowered, and when the level was at its lowest and highest stages, have been prepared to show the relation between the water level in Fresh Pond and in the lines of wells extending outward from it. One of these profiles extending to the north (see Fig. 5) includes wells Cam. 785, 126 ft deep, Arl. 174, 176 ft deep, and Arl. 172, a shallow well. The first two wells were driven into the deep gravel strata beneath clay beds and the water levels in them follow closely changes in storage in Fresh Pond, indicating that there is a hydraulic connection between the pond and the deep gravel beneath the clays north of the pond. The record of well Arl. 172 has been placed on this profile to show that its water level is not affected by fluctuations of stage at Fresh Pond.

In Figure 6 are shown water-level profiles for a pair of wells on a line north of the pond. Well Cam. 796, 34 ft deep, is about 550 ft from the pond and Cam. 802, 53 ft deep, is 1100 ft from it. The former ends in sand, the latter in sand and gravel. Test wells Cam. 795 and Cam. 794 were driven in this line about 2100 and 4000 ft, respectively, from the pond. Both were unsuccessful, ending in clay. In the wells near the pond, Cam. 796 and Cam. 802, the fluctuations follow Fresh Pond, but farther north the impermeable clays prevent flow of ground water between the isolated areas of shallow water-table gravels.

Another line of wells, northwest of Fresh Pond, includes Cam. 797, 798, 263, 799, 264, and Bel. 32 (see Fig. 7). These water levels represent conditions in the shallow sands and gravels near the land surface. The water levels in the first four wells follow the trend of Fresh Pond.

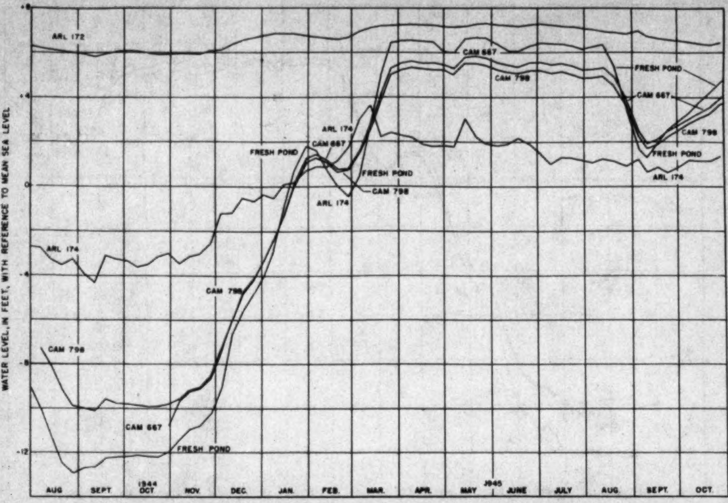

Fig. 4--Hydrographs based on weekly measurements, showing the fluctuations of water levels in Fresh Pond and in wells Arl. 172, Arl. 174, Cam. 667, and Cam. 798

However, the fluctuations are noticeable to a lesser extent at Cam. 264 and Bel. 32. The water level in well Bel. 32 is also affected by changes in stage at nearby Little Pond.

To the south of Fresh Pond, water levels were observed in wells Cam. 801, 819, 820, and Wat. 111 (see Fig. 8). The effect of changes in stage at Fresh Pond can be readily observed in Cam. 801 and Cam. 819 but not definitely on Cam. 820. Well Wat. 111 shows the effect of recharge after heavy pumping stopped at a nearby industrial plant in November 1944, and also the effect of pumping at the same plant after the middle of March 1945.

As a part of the investigation, weekly observations of water levels were made at four wells in the Aberjona Valley, three being situated just north of the Fresh Pond area in Woburn and one in Wilmington. A water-level recorder was installed on one of these wells during December 1944 to obtain a continuous record of water levels. Observations were made to obtain a record of the fluctuation of water levels in nearby glacial deposits not affected by the rise and fall of Fresh Pond. The water level in the four wells shows the same general pattern, controlled chiefly by precipitation. Those wells around Fresh Pond that are affected by changes of stage in the pond often show different patterns from those in the Aberjona Valley. For example, in the Aberjona Valley there was a steady decline in water level from the first part of December 1944 to the middle of February 1945, whereas the water table in the Fresh Pond area rose several feet during this period.

Ground-water yield of Fresh Pond--The purpose of this study was to determine the yield of the glacial deposits surrounding Fresh Pond. Hence it was necessary to measure all inflow and outflow affecting pond stages.

Some of the changes in ground-water storage in the glacial deposits surrounding Fresh Pond have been computed by combining algebraically the figures for inflow to and outflow from the pond, and changes in pondage due to precipitation and evaporation. Inflow to the pond from sources other than those natural to it consists of water passing over an overflow weir on the pipe line from Stony Brook Reservoir to the pumping station, and wash water which is returned to the pond when the filter beds are washed. Controlled outflow consists of pumpage from Fresh Pond to the filter plant, and overflow through the pipe lines to the storm sewer and to Alewife Brook. Actual change in pond storage has been determined from data furnished by the City of Cambridge.

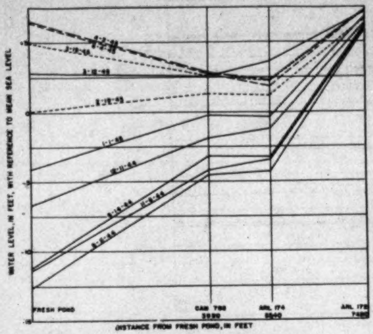

Fig. 5--Profiles showing relation between the water levels in Fresh Pond and in wells Cam. 785 and Arl. 174, ending in deep gravels north of the pond; shallow well Arl. 172 included for comparison

As records for inflow into Fresh Pond from Stony Brook and Hobbs Brook Reservoirs were not available for the entire period of the study, no attempt has been made to compute ground-water inflow or outflow for the complete period. Instead, short periods were used at several different pond stages.

Computations indicate that during the month of July 1944, when the pond was being drawn down, ground water was drawn into the pond from the surrounding sands and gravels in the amount of 286 million gallons during the month, or about 9.2 million gallons a day. In August, when the water level in the pond was still lower, about 246 million gallons were drawn from ground-water storage, representing an average rate of about 7.9 million gallons a day. During the period when the pond was lowest, September 1-9, 1944, and when the hydraulic gradient of the water table was lowest, only about 3.1 million gallons a day were drawn from the formations surrounding the pond.

In March, 1945, when the water level of the pond was rising sharply, the direction of flow was reversed and water began flowing away from the pond into ground-water storage. From this time on during the period when the water level in Fresh Pond was high, considerable water flowed into the sands and gravels near the pond.

These data indicate that water will flow either from or into the sands and gravels surrounding Fresh Pond, depending upon the relation between ground-water levels and pond stage. The water-level profiles (see Fig. 5 to 8) give evidence of storage, particularly in the glacial deposits underlying the area immediately northwest of the pond and in the vicinity of the shore along the west and north perimeters of the pond. South of the pond, ground-water fluctuations related to pond levels were observed in well Cam. 819, situated about 1400 ft from the pond and possibly at the location of well Cam. 820, about 2,300 ft from the pond.

Quality of water

As part of this study analyses were made of ground water from 24 locations in the area covered by this report. Of these, 12 were made of water samples collected from test wells driven by the Massachusetts Department of Public Health, in Arlington, Belmont, and Cambridge. Analyses were made by the Geological Survey of water from 12 locations in Cambridge, Watertown, Winchester, and Woburn. Three analyses of the water in Fresh Pond were made by the State Department of Public Health. Also, from records of the Health Department, quality-of-water data were assembled for several wells in Cambridge.

Quality-of-water data were assembled for this report to indicate the mineral constituents that may be expected in the water in this area and to determine if general movements of ground water in the area around Fresh Pond can be traced by such data.

The hardness of the water in wells around Fresh Pond, based on samples collected in June, July, and August 1944, ranged from 59 to 198 ppm. During these months the water level in Fresh Pond was lower than that in the wells and hence ground water was flowing toward the pond. The hardness of the water in the six wells of the Superior Laundry Company (Cam. 802 to 807) was 198 ppm, or somewhat greater than that of the water in the nearby test wells (Cam. 797, 798, and 799) northwest of Fresh Pond, which ranged in hardness from 134 to 174 ppm.

The analysis of the water in well Cam. 796 evidently indicates local conditions, as its hardness of 59 ppm is much less than that in other wells nearby. Certain other wells, (Cam. 800 and 801) had a hardness of 59 and 60 ppm which is much less than that in Cam. 797, 798, and 799, again showing local conditions in the vicinity of the wells. The wells between Cambridge and Watertown (Cam. 819, 820) have much softer water, as does also well Cam. 801.

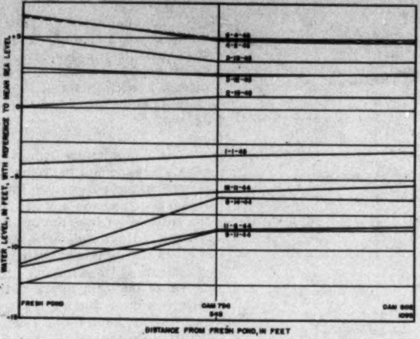

Fig. 6--Profiles showing relation between water levels in Fresh Pond and in wells Cam. 796 and Cam. 802

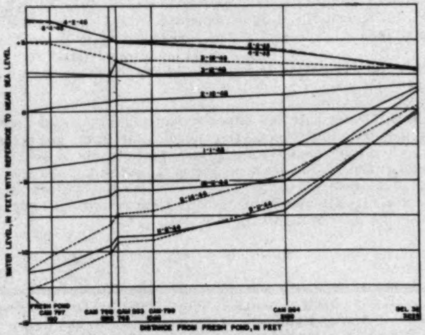

Fig. 7--Profiles showing relation between the water levels in Fresh Pond and in wells ending in shallow aquifers in direction of Little Pond

When well Cam. 669, situated near well Cam. 667 on the golf course near Fresh Pond, was test-pumped in 1931 the water had a hardness of 160 ppm when pumping began. Under continued pumping the hardness rose quickly to 210 ppm and reached a maximum of 260 ppm after 112 hours of pumping at 125 gal/min. This well, drilled 140 ft into a bed of coarse sand, is one of a group of six test wells drilled by the City of Cambridge during 1931 in an attempt to locate a supplementary ground-water supply. The yield obtained was considered inadequate and the water too hard.

Samples of water at three different depths in Fresh Pond were collected at a time when the water level in the pond was low and no water had entered the pond from Stony Brook Reservoir for more than two months. Hence, at this time the water in the pond was supplied from the surrounding ground water and from precipitation. The hardness of the pond water ranged from 65 ppm at the surface to 68 ppm at the bottom. The chloride content of the several well waters ranged from 7.6 to 78 ppm. The chloride content of the pond water was 13 ppm at the surface and 16 ppm at the bottom. There appears to be very little relation, however, between the mineral constituents of the water in the shallow wells immediately surrounding the pond and the pond water. Wells Cam. 796, 800, and 801 have some constituents that approximate those in the pond water, but waters from wells Cam. 797, 798, 799, and Bel. 32, of which Cam. 797 is closest to the pond, seem to have little relation to the pond water.

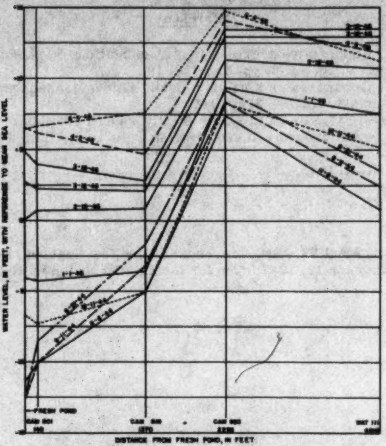

Fig. 8--Profiles showing relation between water levels in Fresh Pond and in wells south of the Pond

The hardness and other mineral constituents of the water in wells in the Fresh Pond area have such a wide range that probably the analyses show only conditions in the immediate vicinity of each well. Hence mineral characteristics of the various waters offer no evidence as to the direction of the movement of ground water.

Summary and Conclusions

The data show that water levels in certain observation wells approximate but lag behind the fluctuations of Fresh Pond, indicating ground-water contributions to the pond when it is at low stage and increases in ground-water storage when it is at high stage. The data further indicate that the largest changes in ground-water storage occur in the glacial deposits underlying the banks west and northwest of the pond.

Thus, recharging the glacial deposits in the vicinity of such ponds could be considered as a means of storing, for later use, a supply of water over and above that which can be stored in the Pond itself. Whether the stored water is later pumped from wells located in the deposits or is pumped directly from the reservoir itself may not make much difference, as in either case the water is available from the pond and the surrounding formations. Pumping from the pond should probably be reserved for situations where the amount of water needed is large. If the amount of water needed is relatively small, it might be advisable to consider pumping from wells instead, inasmuch as wells can be much more easily protected from pollution than can a pond.

Acknowledgments

The writers are indebted to R. K. Hale, Director, Division of Waterways, Massachusetts Department of Public Works; A. D. Weston, Chief Engineer, Division of Sanitary Engineering, Massachusetts Department of Public Health; T. R. Camp, consulting engineer; I. B. Crosby, consulting geologist; T. W. Good and the late G. F. Hooker, of the City of Cambridge, for much assistance in carrying out studies; the Boston Society of Civil Engineers; and the many engineering firms, well owners, and well drillers who gave their hearty cooperation, particularly the Raymond Concrete Pile Co., the Layne-New York Co., and the B. F. Smith Co. The writers also express their appreciation particularly to Messrs. H. B. Kinnison, L. W. Currier, N. E. Chute, and M. L. Brashears, Jr., of the Geological Survey, for furnishing hydrologic and geologic data.

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(Manuscript received April 29, 1948; presented at the New England Meeting, Cambridge, Massachusetts, September 19, 1947; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

WATER-TABLE ELEVATIONS IN SOME PACIFIC COAST BEACHES

J. D. Isaacs and W. N. Bascom

The University of California in cooperation with the Bureau of Ships, United States Navy, has been engaged in field investigations of beach and surf conditions at many of the more important beaches of the West Coast of the United States. One of the first problems encountered in these studies was that of establishing a datum on each beach which would be related to the tidal datum (Mean lower low water) without running long lines of levels from points of known elevation. In an effort to discover a simple way of determining this datum, investigations were initiated, by means of which it was hoped to be able to connect, empirically, the elevation of the ground water in the beach with the predicted tide. Although the original purpose did not prove practicable, considerable interesting data have been obtained. It is the purpose of this paper to present some of these data for the benefit of other investigators.

The method of gathering data was first to make a stadia profile of a beach at right angles to the water edge. Then holes were dug at selected intervals along the profile to depths below the surface of the ground water. At one-hour intervals during one complete tidal cycle the elevation, temperature, and specific gravity of water in the holes, and elevations of uprush and backrush of waves on the beach were observed and recorded. This information then was plotted in profile form to show the relation between the sand and water surfaces and the salinity gradient. Ten beaches were studied in this manner from Carmel, California, to Point Grenville, Washington (see Table 1 for a list of beaches). The slopes of these beaches ranged from 1/10 to 1/150. Precipitation at the various localities ranged to 25 to 100 inches per year.

Table 1--Pacific Coast beaches surveyed for water-table profiles

Name of beach	State
Grenville	Washington
Ocean City	Washington
Leadbetter	Washington
Oysterville	Washington
Clatsop Spit	Oregon
Sand Lake	Oregon
Seabright	California
Seacliff State Park	California
Del Monte Creek	California
Carmel	California

Typical profiles are shown on Figures 1 and 2. The beach at Point Grenville, Washington (see Fig. 1), is relatively narrow and mountains rise abruptly at the backshore. In this instance ground water is flowing from the relatively high backshore, through the beach material into the ocean. Figure 2 shows conditions as existing where sand dunes separate a fresh-water lagoon and the ocean. In this figure water-table profiles are shown as they existed at various times throughout a tidal cycle. In general, the water surface trace is a sloping line which consists of two parts: a seaward section which fluctuates with the tide, and a landward section which remains relatively constant. On a beach on which surf impinges, successive uprushes saturate the sand and give an approximation of still water at that elevation. As the tide advances, the uprushes run higher and higher and raise the area of saturation on the beach face. Soon there is a tendency for salt water to flow into the beach; at the same time the fresh water head is reduced because of the tide. At maximum tide, the trace of the water table at the seaward section becomes a curve which is concave upward. This undoubtedly is because of the relatively slow movement of water in the beach as compared with the change of tide level. As the tide recedes, the curve gradually straightens and then turns downward. On the beaches which were investigated, tides only affected ground water in a narrow area of the beach face. The tide wave might be considered to have entered the beach and to have proceeded through the sand with reduced amplitude corresponding to the duration of tide, permeability of the sand, and distance traveled.

For those conditions where sand dunes exist between the backshore and the beach (see Fig. 2) no rise in the water table under the crest of the dunes, as a result of precipitation was observed, though in some cases there had been continuous precipitation for about a week before gathering these data. This conclusion is contrary to the findings at dunes in certain arid regions where the water table rises in a curve roughly parallel with the surface [LONGWELL, KNOPF, and FLINT, 1939, p. 186].

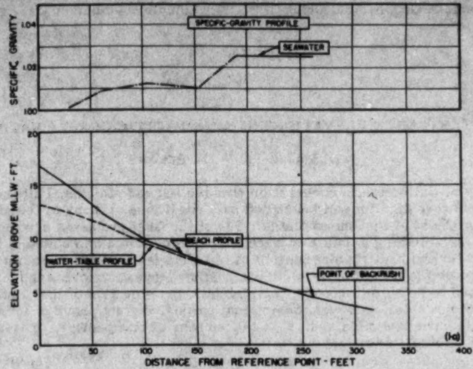

Fig. 1--Profiles of water-table, ground surface, and ground-water specific gravity, Grenville Beach, October 10, 1946

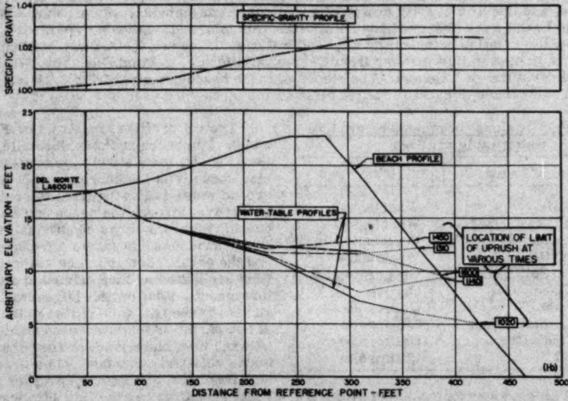

Fig. 2--Profiles of water-table, ground surface, and ground-water specific gravity, Del Monte Creek, December 13, 1946

The specific gravity gradient in all cases was directly related to the water-table profile, that is, a relatively steep water-table profile was accompanied by a relatively steep specific gravity gradient

It is apparent that there is considerable flow of fresh water across ocean beaches. An average rate of flow (measured by dying the water with potassium permanganate and fluorescein) was found to be about one ft/hr.

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(Communicated manuscript received August 6, 1947; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

COMPARATIVE RAIN MEASUREMENTS AND RAIN-GAGE PERFORMANCES ON A STEEP SLOPE ADJACENT TO A PINE STAND

G. L. Hayes and J. Kittredge

Abstract--In this study, measurements from two rows of rain gages, extending outward along contours from the edge of a pine plantation on a steep slope, showed that the 45 foot-high trees influenced the amount of rain reaching the ground for a distance not exceeding 34 ft from the edge of the plantation.

Seven types of rain-gaging installations were tested for accuracy on a windy, 40 per cent slope, including three types of shielding, gages exposed with the orifice horizontal and with orifice tilted parallel to the slope. Best results were obtained with tilted, shielded gages, but tilted, unshielded gages were only slightly less accurate for the conditions encountered. Shielding of untilted gages proved much less effective than tilting unshielded gages, although superior to a conventionally exposed gage, unshielded and with orifice horizontal. It was demonstrated again that the conventional installation is unsuitable for accurate rain gaging on windy slopes.

Introduction--How far from a pine plantation 45 ft high must rain gages be placed to be beyond the influence of the trees; and which of several types of rain-gaging installations will catch the truest sample of the rain falling on a steep slope exposed toward the rain-bearing winds? These two questions arose in connection with a study of rainfall interception by the pines and of runoff and erosion from the pine and from an adjacent grassland area. Before the relations of erosion or runoff could be evaluated it was essential that an accurate measure be obtained of the actual rainfall on the different parts of the area.

To answer the first question, rain gages were placed at measured intervals along first one and then a second line extending on the contour from the edge of the pine stand into the grassland. Rainfall was measured and recorded for two years along the first line and for two and a fraction years along the second.

To investigate the second question seven types of rain-gaging installations were set out on the 40 per cent slope of the grassland, far enough away from the trees to be definitely beyond their influence. Comparative measurements recorded over three- to seven-year periods have been analyzed to determine the relative efficiency of each.

When obstructions are absent from the gaging area, the inaccuracies of rain measurements made on windy slopes arise principally from two sources. The first is the disturbance produced in the rain-bearing wind stream by the presence of the gage. The significant features of this disturbance are an increase in pressure on the windward side of the gage, a decrease in pressure and marked acceleration of wind over the top of the gage, the very area through which the rain must fall to enter the receiver , and eddy currents over and within the orifice [BROOKS, 1938a]. The second source of error is the difference in the angles at which the planes of the orifice and the ground surface intercept the falling rain. Any method or device that is used to improve gaging accuracy should reduce the effects of either one or both of these sources of error.

In this study, the rows of gages which extended perpendicularly from the edge of stand were placed with their orifices parallel to the 40 per cent slope to reduce the second source of error. The effectiveness of this practice has been demonstrated [BROOKS, 1938a; PAGLIUCA, 1934; STOREY and HAMILTON, 1943; STOREY and WILM, 1944]. Nothing further was done to improve the aerodynamics of these gages. The various installations compared for gaging efficiency included four horizontally exposed gages, one of which was unshielded and similar to a standard installation, and one each was equipped with a Nipher shield, an Alter shield, and placed in a pit, treatments which reduce the first source of error only [ALTER, 1937; BROOKS, 1938, 1938 a and b; CODD, 1938; HAYES, 1944; INSTITUTE OF WATER ENGINEERS, 1937; RIESBOL, 1940]. Also included were two of the tilted unshielded gages, a tilted, Nipher-shielded gage [BROOKS, 1934, 1938a; PAGLIUCA, 1934], and trough-type gages [STOREY and HAMILTON, 1943]. The latter two installations reduce both sources of error.

Description of the area and the installations--The study area is located on a southeast-facing, 40 per cent slope in Strawberry Canyon about one-half mile east of the University of California at Berkeley. It is 370 ft above the canyon bottom and 200 to 400 ft below a spur ridge of the Berkeley Hills, to the west and north. It is freely exposed to the rain-bearing winds which characteristically come from the southeast at the beginning of a storm, changing through south to west toward the end. The rain-bearing winds therefore usually move either up-slope or angle upward across the slope. The area is free of obstructions on all but the southwest side where the edge of the 45-foot Canary Pine plantation is shown by Figure 1 to extend up and down the slope in a northwest-southeast direction to the point where it is joined by the lower of the two lines of rain gages extending northeastward from the stand. Below this point the edge of the stand is oriented roughly north and south. During the early stages of a typical storm, the winds usually passed over the area, upslope, approximately parallel to the border of the stand, then gradually changed to cross the area somewhat to leeward of the stand. In occasional storms the wind moved from other directions. The rains were mostly of low intensity and the accompanying winds of low velocity. Both intensities and velocities tended to increase with the amount of rain and in the later parts of the storms. The ground surface was covered with grasses which averaged about six inches high during the rainy season.

The locations of all gages to be discussed are designated by symbols on Figure 1. The first term of the symbol is in all cases either the letter H, indicating that the orifice of the gage was horizontal, or T, indicating that the orifice was tilted parallel to the 40 per cent slope. The second term of the symbol, if a number, indicates that the gage was one of the series which extended from the stand into the gaging area, the number denoting the distance in feet from the stem of a tree in the outermost row of the plantation. If composed of letters the second term designates the type of gage or type of shielding used, as will be described below.

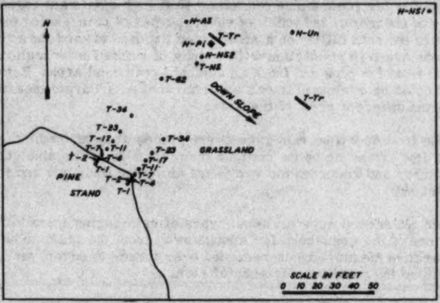

Fig. 1--Locations of gages

The numbered gages occupied the upper of the two rows for the 1939-1940 and 1940-1941 rainy seasons and the lower row from then until January 2, 1944, when this phase of the study was terminated. The gages at one, two, and four feet from the stems were underneath the crowns of the trees. The one at four feet in both rows was beneath the edge of the crown. All were tilted, unshielded, and placed with the orifice approximately 26 inches above the ground. The remaining installations, all of which were placed with the orifices 26 inches above ground unless otherwise noted, were as follows.

T-NS, a standard eight-inch rain gage, tilted with its orifice parallel to the 40 per cent slope and equipped with a Nipher shield with wire screen of the Harvard Blue Hill pattern of 1938, described by BROOKS [1938b]. It remained in the same location 80 ft northeast of the pine stand for the entire seven years of the study. For reasons which will be explained later, this was used as a reference gage with which all others were compared.

H-NS, identical with T-NS except that the orifice and shield were horizontal. From September 1939 to January 1944 it was located 135 ft east of T-NS after it was relocated only five feet northeast of T-NS.

H-Un, a 12-inch, tipping-bucket type recording rain gage which was used for current measurements only. The rain was collected in the measuring tube and measured with a stick in the usual way With vertical sides and the orifice horizontal and unshielded, this most nearly represented a standard installation. It was located 57 ft east of T-NS for the whole seven years of study and also for the four preceding years as a basis for comparisons with gage H-Pi.

H-AS, a standard eight-inch gage exposed on a steel tower with the horizontal orifice eight feet above ground, and equipped with an Alter shield [ALTER, 1937; CODD, 1938]. It was located 24 ft north of T-NS for only the last four years of the study. This is an installation that is being widely tested by the United States Weather Bureau for measuring snow, especially at infrequently visited stations.

T-Tr, the two gages bearing this designation were trough gages four inches wide by ten feet long and equipped with movable slides to permit adjustment of the receiver openings to a horizontal area of 427.2 sq in. Each was installed with its long axis up and down slope and its rim four to five inches above ground and one inch above the clipped grass. The dense cover of dead and living grasses minimized splash from the ground surface into the gages. For comparison with the other gages, the lower trough gage was kept free of over-topping grass by clipping around it from September 1939 to April 4, 1944, after which the grass was permitted to grow over it and the upper gage was kept open for comparison. The upper gage was located 18 ft northeast of T-NS and the lower one 50 ft straight downhill (southeast) from the upper.

H-Pi, a standard eight-inch gage with orifice horizontal and exposed in a pit only slightly larger than the gage with the receiver approximately even with the clipped grass. As with the trough gages, the dense cover of grasses, which were kept from over-topping the gage by clipping around it, were relied upon to prevent splash from the ground surface. As this installation preceded most of the others and was removed when they were installed, it could not be compared directly with any but H-Un for the four years preceding May, 1939, and with T-Tr for the 1938-1939 season. It was located 55 ft northwest of the lower trough gage and 46 ft west of H-Un during the period of comparison.

Procedure--Measurements were made, not at any set time of day, but shortly following the cessation of each storm. Single measurements therefore often included the sum of two or more days of rain whereas in other cases two or more storms were measured on the same day. Single storms varied from as little as 0.01 to more than eight inches of rain.

The catches of the tilted gages were corrected for the reduced horizontal projection of the orifice area by multiplying by the secant of the angle between the plane of the orifice and the horizontal, and the catches of the trough gages were converted to inches depth on horizontal area to make all gage catches comparable.

The total rain caught at the different distances from the pine stand by years and by groups of years which correspond to the upper and lower locations of the row of gages is shown in Table 1.

Table 1--Summary of rain measured by years and groups of years at different distances from the pine stand

Year	Distance from tree stem in feet										No. of storms
	1	2	4	7	11	17	23	34	62	80[a]	
	in	in	in	in	in	in	in	in	in	in	
1939-1940	38.19	28.29	27.09	23.13	25.29	29.87	33.40	35.85	35.68	36.57	38
1940-1941	26.81	22.07	27.23	28.88	33.53	37.77	40.14	42.81	41.48	42.12	59
Upper position subtotal	65.00	50.36	54.32	52.01	58.82	67.64	73.54	78.66	77.16	78.69	97
1941-1942	20.08	22.81	30.80	27.84	31.68	33.32	36.33	36.05	36.34	36.79	48
1942-1943	15.84	19.62	21.64	21.33	23.33	25.19	26.88	27.21	....	28.39	43
1943-1944	2.79	3.55	4.04	4.74	5.43	5.69	5.96	5.87	....	5.96	12
Lower position subtotal	38.71	45.98	56.48	53.91	60.44	64.20	69.17	69.13	....	71.14	103
Grand total	103.71	96.34	110.80	105.92	119.26	131.84	142.71	147.79	....	149.83	200

[a] Tilted and Nipher-shielded reference gage, T-NS.

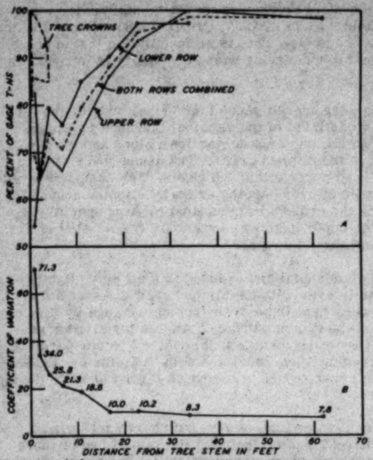

Fig. 2--(A) Catch of tilted, unshielded gages in rows extending from edge of pine stand expressed as percentages of the catch of the tillted, Nipher-shielded gage 80 ft from the stand; (B) coefficients of variation in the rainfall catch at different distances from the pine stand; data from both rows combined

The relation between the catches is more clearly shown, however, by Figure 2A in which the totals for each of the rows separately and for the two rows combined are expressed as percentages of the catches by the reference gage, T-NS. The amount of rain caught did not increase beyond 34 ft from the stand in the upper row, and did not increase materially beyond 23 ft in the lower row. At greater distances, variations in the catch appear to be random and non-significant.

To test this conclusion, the significance of the difference between the means of the catches of adjacent gages from 17 to 62 ft was tested by the technique described by PATERSON [1939, pp. 146-147]. The tests substantiated statistically the conclusion that the rain catches did not change significantly beyond 34 and 23 ft from the stand in the upper and lower rows, respectively, but did increase significantly out to these distances. When the catches of both rows were combined the difference in number and size of storms included in the total for 62 ft made a comparison of means impractical so the differences between regression coefficients were tested. Again, the increase in catch was significant out to 34 ft from the stand and not significant beyond there.

Figure 1 suggests a reasonable explanation as to why the stand influence should extend farther along the upper than the lower row of gages. Because of the change in direction of the stand boundary at the point where it is joined by the lower row, storm winds of southerly direction which would have to cross the stand to reach the 23-foot gage in the upper row, could reach the same position in the lower row without crossing the trees. It was therefore concluded that gages T-34 and T-62 were beyond the influence of the pine stand and their catches were therefore included in the next analysis to determine which kind of installation provided the best measure of the actual rainfall.

The catches of all the different types of installations, except the pit gage are summarized and compared by years and groups of years in Table 2, and for the pit gage and others which were operated concurrently with it, in Table 3. The groups of years in Table 2 represent periods during which the locations of all gages remained unchanged except for the minor moves of gages T-34 and T-62 preceding the 1941-1942 rainy season. It may be noticed that the number of storms included for the same years may differ slightly in Tables 1 and 2. This is because storms for which the record was faulty from one or more gages were eliminated from one analysis but were included in the other analysis which concerned a different combination of gages.

The greatest total catch was recorded for the trough gages followed closely by the tilted and Nipher-shielded. In general, the tilted gages, including the trough type, caught more than the untilted, whether shielded or not. The Nipher-shielded caught more than the unshielded or Alter-shielded if only tilted or untilted gages are compared. The pit gage caught about the same amounts as the untilted, Nipher-shielded gage, H-NS. Although the comparison cannot be made directly, the four-year total caught by H-Pi, Table 3, was 106 per cent of the catch of H-Un, and the seven-year total caught by H-NS (see Table 2), is 105 per cent of the catch of H-Un.

Because of the differences in the number and sizes of storms measured by the different installations the full record from each installation could not be used for comparing totals or means. Regression coefficients, however, permit the use of all the data and their comparison is more instructive for highly correlated records such as these, but the use of regression coefficients requires that the catches of the respective gages be related to a standard. The trough gages, which

Table 2--Total rain measured by the different gaging installations by years and groups of years

Year	Gage							No. of storms
	T-Tr	T-NS	T-34	T-62	H-NS	H-Un	H-AS	
	in	in	in	in	in	in	in	
1939-1940	37.16	37.47	36.73	36.38	34.16	31.68	. . .	41
1940-1941	43.45	41.68	42.39	41.04	38.53	36.50	. . .	57
1941-1942	37.35	36.77	36.04	36.34	35.38	31.90	. . .	47
Three-year subtotal	117.96	115.92	115.16	113.76	108.07	100.08	. . .	145
1942-1943	28.47	28.40	27.21	. . . .	26.80	24.45	25.30	43
1943-1944	23.47[a]	23.14	. . . .	. . . .	21.31[b]	21.13	20.43	38
1944-1945	26.10	27.09	. . . .	. . . .	24.91	24.90	24.09	37
1945-1946	23.74	23.40	. . . .	. . . .	21.10	21.49	20.37	42
Three-year subtotal	73.31	73.63	. . . .	. . . .	67.32	67.52	64.89	117
Seven-year totals	219.74	217.95	. . . .	. . . .	202.19	192.05	. . .	305

[a] The lower trough gage was replaced by the upper after the 27 storm of the 1943-1944 season.

[b] Gage H-NS moved from 146 ft east to five feet northeast of T-NS after the 12 storm of the 1943-1944 season.

Table 3--Comparison of the pit gage with the horizontal, unshielded, and trough gages

Year	Gage			No. of storms
	T-Tr	H-Pi	H-Un	
	in	in	in	
1935-1936	. . .	21.84	21.04	41
1936-1937	. . .	25.58	24.13	47
1937-1938	. . .	30.60	28.71	47
1938-1939	14.73	14.04	13.14	34
Total	. . .	92.06	87.02	169

are aerodynamically the most efficient [STOREY and HAMILTON, 1943], probably caught the truest sample of the actual rainfall. But because the trough gages represent an infrequently used type, and because preliminary tests showed that no significance could be attached to the difference between the catches of these and the tilted, Nipher-shielded gage, the latter was selected as a reference gage.

There remained the questions of whether the large change in location of gage H-NS on January 1, 1944, and the change from the lower to upper trough gage on April 4, 1944 significantly altered their catches. The regressions of H-NS on T-NS for the periods before and after the move proved to differ significantly, but as the relative position of the regression in relation to the others was not changed a single regression for the whole seven years was adopted. This decision was later justified by the standard error of this single regression which proved to be one of the two lowest that were computed (see Table 4).

A test of the differences between the means of the catches of the upper and lower trough gages, using the method of Student [PATERSON, 1939, p. 20], for two periods including 43 storms when both were free of overtopping grass, showed no significant difference between them. Likewise, the regressions of the catches of each gage separately with the catches of gage T-NS proved non-significant. One regression for the whole seven years was therefore used to represent the trough gages.

Discussion of results--Although the main purpose of the rows of gages which extended outward from the edge of the pine stand was to find out how far the stand influenced the amount of rain reaching the ground, other points of interest are illustrated by Figure 2. These include: (1) The large variability in catch of the gages located under the tree crowns. This is illustrated by the wide differences in percentage catch of the gages at one and two feet from the tree stem, ranging from 83.6 to 54.4 per cent of the reference gage catch (see Fig. 2A). (2) The catch averaged higher at four feet, under the edge of the crowns than on either side. Drip from branch tips evidently accounted for this. (3) Variations in the amount of catch, over and above variations due to differences

Table 4--Summary of regression coefficients with their standard errors for the catches of designated gages related to the catch of gage T-NS

Item	Gage							
	T-Tr	T-NS	T-62	T-34	H-NS	H-Un	H-AS	H-Pi
	in	in	in	in	in	in	in	in
Regression coefficient	1.004	1.000	0.989	0.981	0.906	0.878	0.844	0.916[a]
Standard error	±0.0044	. . .	±0.0048	±0.0041	±0.0039	±0.0039	±0.0068	b
Y-intercept	0.0034	0	0.0062	0.0041	0.0155	0.0023	0.0255	0.0098
Basis (No. of storms)	305	305	145	188	305	305	162	169

[a] Computed from the relation of H-Pi to H-Un: H-Pi = 0.0074 + 1.0436 H-Un and H-Un = 0.0023 + 0.8780 T-NS. From these equations the relation of H-Pi to T-NS was computed to be H-Pi = 0.0098 + 0.9162 T-NS.

[b] Standard error of regression of H-Pi in relation to H-Un was ± 0.0046 which is within the range of the others.

in storm magnitude, were high within the rain shadow of the trees and decreased with distance from the stand to an approximately constant value beyond the limits of the stand influence. This is best illustrated by the coefficients of variation (see Fig. 2B) [PATERSON, 1939, p. 23], which include both variations between rows and between storms. As the wind direction and velocity varied from storm to storm the extent and magnitude of the stand influence could likewise be expected to vary. Within the rain shadow of the stand the coefficients of variation decreased from 21.3 per cent at seven feet from the stand to 8.3 per cent at 34 ft, but only from 8.3 to 7.8 per cent between 34 and 62 feet, thus corroborating the former tests in indicating that the stand influence did not extend beyond 34 feet.

The regression coefficients for the different types of gaging installations, together with their standard errors are assembled in Table 4. All differences between regression coefficients that are greater than 0.012 are statistically significant, all smaller differences are non-significant.

Outstanding among the relations is the similarity of performance of and the superiority of the tilted gages over the untilted whether shielded or not. The regressions of the trough gages, the tilted and Nipher-shielded, and the tilted, unshielded form a close group, and their coefficients exceed those for the best horizontally exposed gages, H-Pi and H-NS, by from 7.1 to 10.8 per cent.

Shielding was less effective than tilting for improving the catch. Shielding untilted gages, as represented by the horizontal, Nipher-shielded, and pit gages increased the coefficients over the horizontal unshielded gage by only 3.2 and 4.4 per cent, respectively. Shielding a tilted gage, T-NS, improved the regression coefficient over the tilted unshielded gages, T-34 and T-62, by only 1.9 and 1.1 per cent, respectively, the latter of which is not statistically significant and the former is of dubious practical significance.

The lowest coefficient of all is that for the horizontally exposed, Alter-shielded gage which was less than the standard installation, H-Un, by 3.9 per cent and less than the tilted gages by from 14.0 (T-34) to 15.9 (T-Tr) per cent. The reason for this poor showing is undoubtedly the greater wind velocities at eight feet above ground than at two feet which more than counter-balanced any beneficial influence of the shield. The conclusion is indicated that, for best results, this installation, which was designed for use in deep snow regions, should be used on level ground.

Conclusions--(1) The presence of a 45-foot high pine stand with its border roughly parallel to the prevailing direction of the rain-bearing winds decreased the amount of rainfall at gage height for a distance not exceeding 34 ft from the trees or 3/4 the mean height of the stand.

(2) The variability in catch from storm to storm, over and above the variations associated with differences in storm magnitude, was greater within than beyond the rain shadow of the stand.

(3) The trough-type rain gage, and a standard type tilted and equipped with a Nipher-shield provided the best measures of true rainfall.

(4) Standard gages, tilted with their orifices parallel to the slope, but unshielded, caught only a little less rain than the trough and tilted, Nipher-shielded gages. This does not mean that shields may not be justified for windier mountain slopes or crests having less cover.

(5) The Nipher shield increased the catch of a vertical standard gage but not as much as tilting without the shield.

(6) The marked inferiority of the horizontally exposed in contrast to the tilted gages once more emphasizes the conclusions of BROOKS [1938a], PAGLIUCA [1934], STOREY and HAMILTON [1943], and STOREY and WILM [1944] that for accurate rain gaging on sloping ground the gage orifice should be parallel to the ground surface.

(7) The Alter shield did not compensate for the aerodynamic disadvantage of raising the gage to eight feet above ground. Such poor performance with rain suggests that the catch of the more-difficult-to-gage snow, is even less likely to be dependable.

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(Manuscript received February 19, 1948; presented at the West Coast Meeting, Berkeley, California, February 13, 1948; open for formal discussion until September 1, 1949.)

Transactions, American Geophysical Union Volume 30, Number 2 April 1949

DISCUSSION OF

"UNSTEADY FLOW IN OPEN CHANNELS" by H. PUTMAN

[Trans., v. 29, pp. 227-232, 1948]

Pin-Nam Lin (Iowa Institute of Hydraulic Research, Iowa City, Iowa)--Two more methods of deriving the equations of characteristics--The writer should like to call attention to two more methods of deriving the equations of characteristics that do not require such mathematical background as was set forth in Massau's work, though the underlying principle is still the same. (1) To begin with, let the partial derivatives be interpreted as follows: $\partial z/\partial x$ is the instantaneous water-surface slope; $\partial y/\partial t$ is the time rate of change of depth at some fixed section; $\partial V/\partial x$ is the instantaneous rate of change of velocity with respect to the distance, positive $\partial V/\partial x$ indicating a velocity that is increasing downstream, and $\partial V/\partial t$ is the time rate of change of velocity at a fixed section.

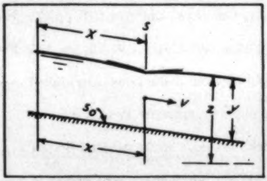


Fig. 1--Schematic sketch of a minute wave, discontinuous surface slopes

Keeping these physical interpretations in mind, one can see that in a general case of unsteady flow, any of these derivatives may be discontinuous, owing to artificial or natural causes of disturbance. For instance, if one considers the situation at section S (see Fig. 1), one can see that the instantaneous slope $\partial z/\partial x$ is discontinuous because two distinct values of $\partial z/\partial x$ exist at the same point. Since neither of the surface slopes bears any definite relationship to the other, $\partial z/\partial x$ must be of an indeterminate form either $0/0$ or ∞/∞. As ∞/∞ is physically impossible, one has $\partial z/\partial x = 0/0$ which will amount to

$$\partial y/\partial x = 0/0 \quad \ldots\ldots\ldots\ldots\ldots (1)$$

because $\partial z/\partial x = s_0 + \partial y/\partial x$. Similar reasoning shall lead to

$$\partial y/\partial t = 0/0 \quad \ldots\ldots\ldots\ldots\ldots (2)$$

$$\partial V/\partial x = 0/0 \quad \ldots\ldots\ldots\ldots\ldots (3)$$

$$\partial V/\partial t = 0/0 \quad \ldots\ldots\ldots\ldots\ldots (4)$$

Now regarding these four derivatives as unknowns, one has four equations to be solved for them.

$$\partial y/\partial x + (V/g)(\partial V/\partial x) + (\partial V/\partial t)(1/g) = s_0 - s_f$$

where $s_f = V^2/C^2R$, Chezy; or $n^2V^2/2.21R^{4/3}$, Manning;

$$\partial y/\partial t + V(\partial y/\partial x) + y(\partial V/\partial x) = 0$$

$$(\partial y/\partial t)\, dt + (\partial y/\partial x)\, dX = dy$$

$$(\partial V/\partial t)\, dt + (\partial V/\partial x)\, dX = dV$$

Algebraic operations will readily show that every one of the conditions, (1), (2), (3), and (4), should independently yield the same equations of characteristics. For example, let us take $\partial y/\partial x = 0/0$. The method of determinants gives

$$\partial y/\partial x = \left[-y\,(s_0 - s_f) + (y/g)(dV/dt) - (V/g)(dy/dt) + (1/g)(dy/dt)(dX/dt)\right]/\Delta = 0/0$$

where $\Delta = (1/g)(dX/dt)^2 - (2V/g)(dX/dt) + V^2/g - y$.

$\Delta = 0$ yields

$$dX = (V \pm \sqrt{gy})\,dt \qquad (5)$$

Setting the numerator equal to zero and keeping (5) in view, one will arrive at the expression $d(V/g \pm 2\sqrt{y/g}) = (s_0 - s_f)\,dt$.

(2) The second method consists in reducing the partial derivatives to total derivatives. Inspection of the equations of motion and continuity suggests that they be added up to bring all partial derivatives together. Dimensional considerations, however, require that the second equation be multiplied by a quantity of dimensions T/L, before it is added to the first. Let this quantity be ϕ. One has then, on adding and rearranging terms,

$$(1/g)\left[\partial V/\partial t + (\partial V/\partial x)(V + \phi gy)\right] + \phi\left[\partial y/\partial t + (\partial y/\partial x)(V + 1/\phi)\right] = s_0 - s_f$$

Should the quantity in the braces be total derivatives, one ought to have $V + \phi gy = V + 1/\phi$; that is, $\phi = \pm 1/\sqrt{gy}$. Therefore

$$(1/g)\left[\partial V/\partial t + (\partial V/\partial x)(V \pm \sqrt{gy})\right] \pm (1/\sqrt{gy})\left[\partial y/\partial t + (\partial y/\partial x)(V \pm \sqrt{gy})\right] = s_0 - s_f$$

Now if, as depicted in Figure 1, X is defined as the distance to a minute wave, open channel hydraulics has well established that

$$dX/dt = V \pm \sqrt{gy}$$

Substituting from this expression, one can have

$$(1/g)\left[\partial V/\partial t + (\partial V/\partial x)(dX/dt)\right] \pm 1/\sqrt{gy}\left[\partial y/\partial t + (\partial y/\partial x)(dX/dt)\right] = s_0 - s_f$$

that is,

$$dV/g \pm dy/\sqrt{gy} = (s_0 - s_f)\,dt$$

or

$$d(V/g \pm 2\sqrt{y/g}) = (s_0 - s_f)dt$$

According to these derivations, the writer is inclined to believe that the sign in (8) of the author's paper should be reversed, namely that the sign in (7) should correspond to the same in (8).

The equation of steady flow--It might be interesting to show here that the equation of backwater curves in the case of steady flow can be deduced from the equations of characteristics, namely

$$dX/dt = V \pm \sqrt{gy}$$

and

$$d(V/g \pm 2\sqrt{y/g}) = (s_0 - s_f)\,dt$$

For steady flow, $Vdy = -ydV$ (the equation of continuity), one can write

$$s_0 - s_f = (1/g)(dV/dt) \pm (1/\sqrt{gy})(dy/dt) = \left[-(V/gy) \pm (1/\sqrt{gy})\right](dy/dt)$$

$$= (1/gy)(-V \pm \sqrt{gy})(V \pm \sqrt{gy})(dy/dX) = (1 - V^2/gy)(dy/dX)$$

$$dX/dy = (1 - V^2/gy)/(s_0 - s_f)$$

which is the equation of backwater curves in prismatic rectangular channels.

A graphical scheme of solving the characteristics--The following is the result of a primary investigation stimulated by the author's statement in his introductory paragraph, "The application of Massau's methods in combination with modern computational procedures should facilitate the study of many problems that have so far defied satisfactory solution."

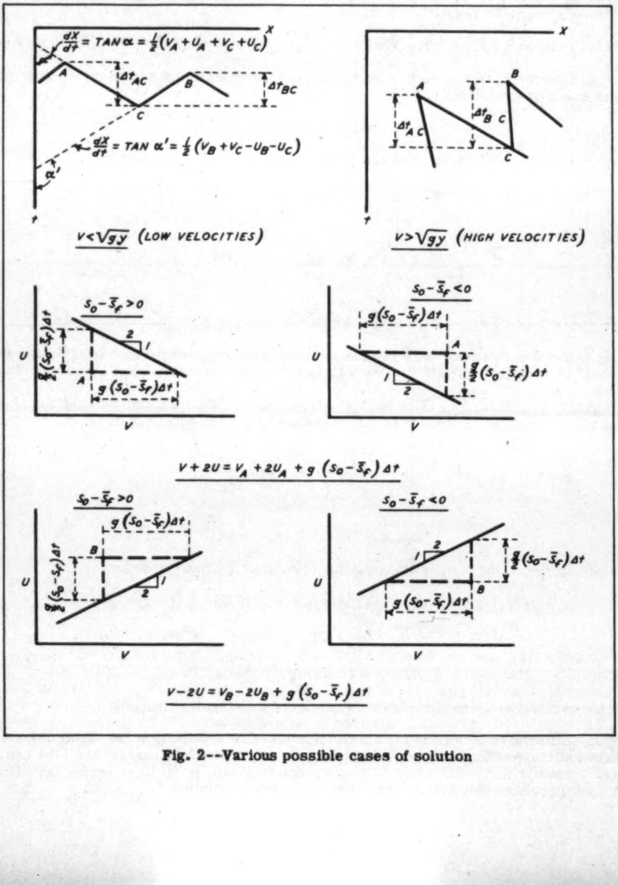
One of the essential steps in this scheme is, after French writers, the introduction of the notation $U = \sqrt{gy}$. In terms of this notation, the equations of characteristics may be written as $dX/dt = V \pm U$, and $d(V \pm 2U) = g(s_0 - s_f)\,dt$. If the whole channel is divided into a number of sufficiently short stretches, s_f would not vary too much. Integration shall then give

$$V + 2U \approx V_A + 2U_A + g(s_0 - \bar{s}_f)\Delta t$$

and

$$V - 2U \approx V_B - 2U_B + g(s_0 - \bar{s}_f)\Delta t$$

where A and B are arbitrary points up- and downstream, respectively; y, V, and t at these points being known; and $\bar{s}_f$ is taken arbitrarily as the arithmetic mean of s_f at ends of the stretch. Evidently these equations are approximately straight lines, when plotted on a V-U plane. Various possible cases are shown in Figure 2.



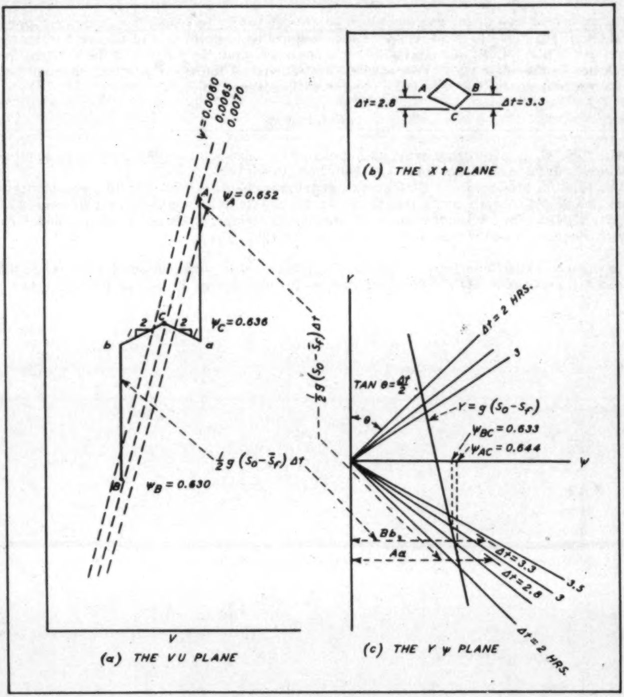
$\frac{dX}{dt} = TAN\,\alpha = \frac{1}{2}(V_A + U_A + V_C + U_C)$
X
A
B
C
Δt_{AC}
Δt_{BC}
$\frac{dX}{dt} = TAN\,\alpha' = \frac{1}{2}(V_B + V_C - U_B - U_C)$
α
α'
t
$V < \sqrt{gy}$ (LOW VELOCITIES)
Δt_{AC}
Δt_{BC}
$V > \sqrt{gy}$ (HIGH VELOCITIES)
$s_0 - \bar{s}_f > 0$
$\frac{g}{2}(s_0 - \bar{s}_f)\Delta t$
$g(s_0 - \bar{s}_f)\Delta t$
2
1
U
V
$s_0 - \bar{s}_f < 0$
$V + 2U = V_A + 2U_A + g(s_0 - \bar{s}_f)\Delta t$
$V - 2U = V_B - 2U_B + g(s_0 - \bar{s}_f)\Delta t$

Fig. 2--Various possible cases of solution

The next step is a graphical evaluation of the quantity $(g/2)(s_0 - \bar{s}_f)\Delta t$. Let $Y = g(s_0 - s_f)$, Then using the Manning formula, we have

$$Y = gs_0 - (n^2g^{7/3}/2.21)(V^2/U^{8/3}) = gs_0 - k\psi$$

where $\psi = V^2/U^{8/3}$. This equation is a straight line on the Y-ψ plane (see Fig. 3c). By superimposing a group of radiating lines making an angle with the Y-axis equal to $\tan^{-1}(\Delta t/2)$, the quantity $(g/2)(s_0 - s_f)\Delta t$ can be readily evaluated, as indicated in Figure 3c.



A
B
C
$\Delta t = 2.8$
$\Delta t = 3.3$
(b) THE X t PLANE
$\psi = 0.0060$
0.0065
0.0070
A
$\psi_A = 0.652$
C
2
2
1
1
b
a
$\psi_C = 0.636$
$\frac{1}{2}g(s_0 - \bar{s}_f)\Delta t$
$\frac{1}{2}g(s_0 - \bar{s}_f)\Delta t$
B
$\psi_B = 0.630$
V
(a) THE VU PLANE
Y
$\Delta t = 2$ HRS.
3
TAN $\theta = \frac{\Delta t}{2}$
θ
$Y = g(s_0 - s_f)$
$\psi_{BC} = 0.633$
$\psi_{AC} = 0.644$
ψ
Bb
Aa
$\Delta t = 3.3$
3.5
$\Delta t = 2.8$
3
$\Delta t = 2$ HRS.
(c) THE Y ψ PLANE

Fig. 3--Values of $g/2\,(s_0 + \bar{s}_f)\Delta t$

The last step is to superimpose a series of ψ = constant curves on the V-U plane. To illustrate this scheme, let us take a simple example of tranquil flow; given U, V, and t at A and B, required to find U, V, and t at C. This may be accomplished in the following steps: (1) Plot A and B on the x-t and V-U planes and read off $\psi_A = 0.652$, $\psi_B = 0.630$ (see Fig. 3). (2) Tentatively lay off AC and BC to intersect at C; determine $\Delta t_{AC} = 2.8$ hours and Δt_{BC} = three hours. (3) Assume $\psi_C = 0.636$ and obtain the mean values $\psi_{BC} = 0.633$ and $\psi_{AC} = 0.644$. (4) Obtain the values of $(g/2)(s_0 - \bar{s}_f)\Delta t$ in the manner shown in Figure 3c. (5) Determine C in the manner shown in Figure 3a. This step is an application of the results of Figure 2. If the value of ψ_C so determined checks approximately with that assumed, the position of C is considered all right, otherwise another value of ψ_C should be assumed and the process repeated.

Henri J. Putman (6 rue du Nord, Le Zoute, Belgium, Author's closure)--In the author's mind "modern computation methods" involves the use of calculating machines. The principles of nomography were given also by MASSAU [1889], starting on page 134

The system of two partial derivative equations of the first order (1) and (2) can be reduced in the form of one equation of the second order. Massau gave also the graphical solution of this equation and shows at this occasion how to solve the equations of the characteristics, first by successive approximation and second graphically [MASSAU, 1900, p. 542]

The values of $s_0 - s_f$ for various given points can easily be found in an earlier paper [PUTMAN, 1946, Fig. 1]. To find $s_0 - s_f$ corresponding to point Q_{20} in Figure 1 of that paper, we have: $SY = b = I/C^2R$, and tan (Q0X) represents V^2, thus the portion of the vertical through Y, which lies in the angle (Q0X) represents V^2/C^2R, and as the whole vertical segment between 0X and Y represents s_0, the remaining portion will represent $s_0 - s_f$

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Corrigenda--The following corrections are noted in this paper as published: (1) p. 229, fourth line, read $\pm$ instead of $\mp$; (2) p. 229, sixth line, read "the + sign in (8) and the - sign in (7) to the - sign in $(\bar{8})$."

REVIEWS, ABSTRACTS, AND BIBLIOGRAPHY

REVIEWS AND ABSTRACTS

BAESCHLIN, C. F., Lehrbuch der Geodäsie, Orell Fussli Verlag, Zurich, xx + 892 pp., 1948.

The word "geodesy" in many modern languages covers more ground than it does in English, in which it may mean little more than land surveying. This work is definitely a treatise on higher geodesy, a worthy successor to Helmert's monumental treatise of more than 60 years ago, "Die mathematischen und physikalischen Theorieen der höheren Geodäsie." Its two parts correspond roughly to Helmert's two volumes.

A successor to Helmert's great treatise is badly needed. After all, in more than 60 years observations have accumulated, observations of the kind that Helmert knew and observations of new kinds. New methods have been devised, new theories propounded.

Perhaps Baeschlin's work does not go as fully into the physical aspects as does Helmert's work did but is is definitely a treatise on mathematical geodesy. Helmert did not discuss the practical side of geodesy, the manipulation of instruments, and the technique of least-squares adjustments. Baeschlin gives some space to three modern instruments. This is the table of contents:

Part I: Mathematical preliminaries--Chapter I, The terrestrial ellipsoid; Chapter II, Triangulation on the ellipsoid and related questions; Chapter III, The geodesic on the ellipsoid of revolution; Chapter IV, Differential formulas for the geodesic; Chapter V, Geodetic triangles and arc measurements; Chapter VI, Elements of map projection; Chapter VII, Geometrical methods for determining the geoid; General theory of isostasy; Appendix, Tables.

Part II: Geodesy based on the theory of the potential--Chapter VIII, The gravity field of the Earth; Chapter IX, The gravity field on an ellipsoid of revolution; Chapter X, Figures of equilibrium of rotating fluids; Chapter XI, The measurement of gravity; Chapter XII, The reduction of observed values of gravity to the geoid; Chapter XIII, Determination of the geoid and of deflections of the vertical from gravity observations; Chapter XIV, The reduction of spirit leveling; Chapter XV, Further developments of the determination of the geoid by means of astronomical leveling; Chapter XVI, The measurement of gravity gradients; Chapter XVII, Oscillations of the terrestrial pole; Appendix, Bibliography and index of names and subjects.

The following chapters deal with recent instrumental developments: Chapter XI, The Vening Meinesz apparatus and the hypothetical pendulum and the Holweck-Lejay elastic pendulum; Chapter XVI, The Eötvös torsion balance. Chapter XVII gives an account of the International Latitude Service.

Even this bulky volume does not exhaust so large a theme. There is, of course, room for individual differences of opinion as to the emphasis on any given aspect of the subject, as indicated by the allotted space, and as to the inclusion or omission of this or that topic. In some portions the mathematical treatment is fairly detailed, in others, because of limitations of space, the author can do no more than set down the results and refer the reader to original sources.

The print is excellent and the numerous formulas are admirably clear. The book can be recommended without reservations to readers having sufficient mathematical training to profit by studying it.

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Transactions, American Geophysical Union Volume 30, Number 2 April 1949

NOTES AND PERSONALIA

NOTES

Corrigenda--The following corrections are noted in the Transactions:

Vol. 29, June 1948, in the paper by Carroll F. Merriam, p. 386, the second sentence in the first paragraph should read, "New wells are being added by the United States Geological Survey in Pennsylvania and so as time goes on the determination of weekly index value becomes less subject to the variations caused by individual wells and freak observations."

Vol. 29, December 1948, in the paper by Ralph W. Powell, p. 882, about the middle of the fourth line in the paragraph headed Development (and in the correction carried on this in the February 1949 issue, p. 149), the equation should be $C^2k = V^{(2 - p)}R^{\beta}$; also in this paper, p. 885, the fifth column heading should have F underscored, for the Froude Number.

Vol. 29, December 1948, in the Contents on the back cover, under Papers in Meteorology, the last one, The Physics of Thunderstorms, should not be indented under Symposium on Long-Range Weather Problems, as it is not a part of the Symposium.

Vol. 30, February 1949, in the paper by J. A. Putnam and J. W. Johnson on p. 67, the second author's name was misspelled.

Vol. 30, February 1949, in the paper by Jacob E. Gair, p. 117, Figure 6 is upside down.

Handbook of fellowships prepared by UNESCO--Study Abroad, a handbook listing more than 10,000 opportunities for international study in 166 fields in 27 countries, including the United States, has been completed by UNESCO. Copies may be secured in the United States through the Columbia University Press, New York 27, N. Y.

French Greenland expedition--An expedition sponsored by the French Academy of Sciences will set forth in the spring of 1949 under the leadership of Paul Emile Victor to explore the Greenland icecap and related phenomena. The expedition intends to explore the possibility of a permanent high-pressure area over Greenland, and make seismic and gravity determinations of the thickness of the ice. It is anticipated that the expedition will continue for two years, occupying a site at about 71° N, 40° W, near the Eismitte Station of the Wegener expedition in 1930-1931.

Opportunities for Americans for research and visiting professorships in Burma and New Zealand--The Department of State has announced the availability of funds under the Fulbright Act for research work and for visiting professorships in various scientific fields in Burma and New Zealand. Interested persons should write to the Conference Board of Associated Research Councils, 2101 Constitution Avenue, Washington 25, D. C. which has been designated by the Department of State to aid in this work.

New German meteorological publication--A new annual German publication on meteorology entitled "Wissenschaftliche Arbeiten des deutschen meteorologischen Dienstes im franzosischen Besatzungsgebiet" is now being published under the direction of Wilhelm Peppler by the Karl F. Haug Verlag, Saulgau, Württemberg. Volume 1 for 1947 has been received in the United States.

Président Théodore Tissier resumes scientific operations--After serving with the Free French Forces from 1940 until 1946, the research vessel Président Théodore Tissier of the Marine Nationale Français was reconditioned at Nantes, and sailed in August 1948 on a three-month oceanographic cruise of the Atlantic, North Sea, and Channel fishing banks. In 1949 she is scheduled to explore the fishing grounds of Newfoundland, Greenland, and the Antilles.

Vacancies for climatological work--The Environmental Protection Section of the Office of The Quartermaster General is inviting applications from persons with basic training in meteorology, geography, or statistics on analysis of climatological and environmental data in grades ranging from P2 to P6.

Hydraulic Conference at Iowa City--The Fourth Hydraulics Conference of the Iowa Institute of Hydraulic Research will be held at the University of Iowa June 12 to 15, 1949. Inquiries concerning the Conference should be addressed to the Director of the Institute.

High-altitude meteorological research station in the Himalayas--A survey is being made by Indian meteorologists of the western Himalaya ranges to select the location for a high-altitude meteorological research station that has recently been authorized by the Government of India. $900,000 have been appropriated to construct and equip this station.

Shell fellowship program--The Shell Union Soil Corporation has announced the availability of 44 fellowships for graduate study in various physical sciences at 36 colleges and universities for 1949-1950. The scholarships include tuition, a stipend of $1200, and an allotment to the University of $300 for equipment and related expenses. Persons who are interested should write to the Shell Fellowship Committee, 50 W. 50th Street, New York 20, N. Y.

Change in the journal, Terrestrial Magnetism and Atmospheric Electricity--The international quarterly journal, Terrestrial Magnetism and Atmospheric Electricity which has been published without interruption since 1896, has been renamed Journal of Geophysical Research, retaining its international flavor, and deliberately broadening its field in line with a general trend of the last two decades. The Editorial office remains at 5241 Broad Branch Road, N. W., Washington 15, D. C. Merle A. Tuve succeeds John A. Fleming as editor.

Jet Propulsion Fellowships offered--The Daniel and Florence Guggenheim Foundation has announced the availability of six $2000 Jet Propulsion Fellowships for two years of advanced training in this field at California Institute of Technology or at Princeton University. Interested persons should write to the Foundation at 120 Broadway, New York 5, N. Y.

Conference on science at Wellesley College--A three-day conference on science, emphasizing both the biological and physical aspects of energy was held March 16-18, 1949, marking the 75th Anniversary of Wellesley College. James B. Conant, president of Harvard University, opened the conference with an address on Science and Common Sense. The speakers included internationally known scientists, with subjects ranging from biological energy, photosynthesis, atomic energy to the energy of stars.

PERSONALIA

Fred Anderegg, until recently in charge of scientific photographic work in the Department of Geology of Princeton University, has been named Supervisor of Photographic Services at the University of Michigan.

Werner A. Baum of the Department of Geography, University of Maryland, has been appointed Associate Professor of Meteorology at Florida State University.

Frederic Darling, Mathematician of the United States Coast and Geodetic Survey since 1919, retired from active duty March 31, 1949.

Beno Gutenberg has been elected a Foreign Member of the Royal Swedish Academy.

Frank B. Jewett, formerly head of The Bell Telephone Laboratories, has been named Chairman of the Building Research Advisory Board of the National Research Council.

Serge Korff of the Department of Physics, New York University, is spending several months in Europe on cosmic ray studies. He expects to visit Denmark, England, France, Italy, Sweden, and Switzerland collating data secured at various stations.

William G. Metcalf and John Holmes of the Woods Hole Oceanographic Institution spent February and March 1949 in oceanographic research at the Arctic Research Laboratory, Point Barrow, Alaska, under the joint sponsorship of the Hydrographic Office and the Office of Naval Research.

Jerome Namias, Chief of the Extended Forecast Section, United States Weather Bureau, left Washington in February 1949, for an extended visit to several of the European weather services. He will spend about three months in Sweden, giving a series of talks on long-range forecasting and conferring with C.-G. Rossby and others on various research problems, and will later visit Belgium, France, Germany, and several other European countries. He will return to the States in October 1949.

Donald W. Pritchard of the Naval Electronics Laboratory at San Diego, has been appointed to head the Chesapeake Bay Institute of Johns Hopkins University.

Thorndike Saville, Dean of the College of Engineering of New York University, has been appointed a member of the National Advisory Health Council.

Le Roy A. Woollett, Jr. is now employed by the Petroleum Engineering Department of the Magnolia Petroleum Company in Brownfield, Texas.

A. Perry Banta, Associate Professor of Sanitary Engineering at California Institute of Technology, died January 23, 1949 at the age of 44. He was born in New York, N. Y., received his baccalaureate degree from Stanford University in 1926, and his Master of Science degree from California Institute of Technology in 1928. From 1927 to 1941 he served as a sanitary engineer for Los Angeles County, and as a Lieutenant Colonel during World War II, following which he took up his duties which continued until his death.

Rollin T. Chamberlin, Professor of Geology at the University of Chicago, died March 6, 1948 at the age of 66. He received his baccalaureate degree in 1903, and his Doctor of Philosophy in 1907, both from the University of Chicago. He spent the first years of his professional career in China and Brazil, and in 1912 entered his work at the University of Chicago as an instructor in geology, in which work he advanced steadily, becoming professor in 1923. He had been a member of AGU since 1932.

Robert W. Clark, Chief Geologist for the Western Gulf Oil Company, died recently. He was born in West Branch, Michigan August 17, 1884 and received his baccalaureate degree in 1908 and his degree of Doctor of Philosophy in 1925 from the University of Michigan, with special work in 1908 and 1909 at the University of Munich and Heidelberg University. From 1909 to 1916 he taught geology at the University of Michigan, following which he entered the field of petroleum geology, working extensively in Oklahoma and California. In 1938 he entered the service of the Western Gulf Oil Company.

Alexander L. du Toit, Consulting Geologist of Cape Town, South Africa died in February 1948 at the age of 69. He was born in Cape Town, and received his baccalaureate and Doctor of Science degrees from the University of the Cape of Good Hope, with further study at the Royal Technical College in Glasgow. Until 1927, he was engaged in geological and irrigation work for the Government of the Union of South Africa, and was subsequently engaged on consulting work.

Francis P. Gilbert, Hydraulic Engineer with the United States Engineer Office in Providence, Rhode Island died January 17, 1949 at the age of 57. He was born in Athol, Massachusetts and received his baccalaureate degree from Massachusetts Institute of Technology in 1914. In the early years of his career, he served in various sanitary and hydraulic engineering activities in the East and in World War I. Following the War, he engaged in sanitary and malarial control work in various parts of the country, and entered Federal Service in 1934 on hydraulic and hydrologic work.

Harry L. Haehl, Consulting Engineer engaged in irrigation and water supply work, died January 29, 1949 at the age of 72. He was born in Cloverdale, California. In 1901 he received his baccalaureate degree, and in 1902, his Master of Science degree, both from Stanford University, and since that time had engaged in private practice as a consulting engineer.

Samuel H. McCrory, recently returned from a United Nations mission to the Near East on flood control and water supply, died February 18, 1949. He was born in Iowa City, Iowa, May 5, 1879 and received his baccalaureate degree from the University of Iowa in 1904. The first two years of his professional life were spent in the practice of engineering in northwestern Iowa, following which he entered Federal service as a drainage engineer. In the development of this work, his responsibilities grew, and in 1931, he became Chief of the Bureau of Agricultural Engineering. When this bureau was merged with others as the Bureau of Agricultural Chemistry and Engineering in 1939, he became Assistant Chief of the new Bureau. During World War II, he was director of the Hemp Division for the Commodity Credit Corporation, which work he continued until he retired in 1946. He had been a member of AGU since 1931.

Charles S. Slichter, Dean Emeritus of the Graduate School of the University of Wisconsin, died October 4, 1946. He was born in St. Paul, Minnesota in 1864, and received his baccalaureate degree in 1885 and his Master of Arts degree in 1886 from Northwestern University. He began his work at the University of Wisconsin in 1886 as Instructor of Mathematics and became Dean of the Graduate School in 1920, retiring in 1935. The study of the movement of ground water was one of his great contributions to hydrology. He had been a member of AGU since 1930.

Henry R. Stevens, Consultant on power development for the Bonneville Power Administration, died September 25, 1948 at the age of 71. He was born in Syracuse, New York, and received his baccalaureate degree in electrical engineering from Syracuse University. He devoted much of his career to the problems of hydroelectric development, and at one time acted as advisory engineer on the construction of hydroelectric projects in Japan. Since 1940 he had been connected with the Bonneville Power Administration.

Joseph W. Thomas, Jr., Head Observer for the Stanolind Oil and Gas Company on seismic and other geophysical exploration work, died recently. He was born in Cheyenne, Wyoming on July 9, 1910. He attended the University of California at Los Angeles from 1931 to 1933, and entered the service of the Stanolind Oil and Gas Company in 1936.

Bailey Willis, Professor Emeritus of Geology at Stanford University, died February 19, 1949 at the age of 91. He was born at Cornwall, New York and received the M. E. degree from Columbia University in 1878. He began his career as a special agent on the Tenth Census, following which he spent three years on the Northern Transcontinental Survey as a geologist. From 1884 to 1916 he was in the Service of the United States Geological Survey, and until 1922 was Professor of Geology at Stanford University. His scientific studies continued, however, up to the time of his death. One of his more popularly known achievements was his forecasting of the Santa Barbara earthquake in 1925. He had been a member of AGU since 1921.

Henry G. Wrocklage, retired member of the seismologic staff of United States Coast and Geodetic Survey, died December 10, 1948 at the age of 70. He was born in New York, N. Y., and from 1916 to 1920 attended the University of California. He entered the service of the United States Coast and Geodetic Survey in 1922, and continued until his retirement in 1947. He had been a member of AGU since 1934.

UNION, SECTION, AND COMMITTEE ACTIVITIES

Francis W. Reichelderfer designated Vice-President of AGU

Francis W. Reichelderfer of the United States Weather Bureau has been designated by the Executive Committee of AGU to fill the unexpired term of Harald U. Sverdrup who has resigned because of the resumption of his residence in Norway. This designation is in accordance with Article 5 of the Statutes.

Report of Committee of Tellers on the change in By-Laws

Earl Ingerson, Chairman
Wilbur S. Burbank
Frank L. Culley
Richard H. Fleming
Woodrow C. Jacobs
David G. Knapp
Ray K. Linsley, Jr.
Mary C. Rabbitt
Francis G. Wells

The Committee met on March 11, 1949, and tabulated the ballots on the proposed amendment to the By-Laws to raise maximum dues for individual members from $5.00 to $7.00 per year. The results are as follows:

Total number of ballots received			2003
In favor of amendment		1432	
Opposed to amendment		499	
Rejected ballots		72	
Ballots without identification	14		
Ballots from non-citizens of U. S.	26		
Blank ballots	2		
New members who have paid no dues	3		
Ballots mailed after March 5	27		

Twenty-Third Regional Meeting held in Los Angeles

The Twenty-Third Regional Meeting was held in the Physics-Biology Building of the University of California at Los Angeles, February 4 and 5, 1949. There was one session held jointly in Hydrology and Meteorology, three sessions in Hydrology, and three sessions in Meteorology, including a two-session Symposium on the Physics of the Upper Atmosphere. There were 36 papers on the program, many of which will be published in the Transactions. More than 200 were in attendance.

Inter-American Conference on Conservation of Renewable Natural Resources

An Inter-American Conference on Renewable Resources was held in Denver, Colorado, September 7-20, 1948. President Bucher designated S. W. Lohman of the United States Geological Survey as AGU representative. Various Federal agencies participated, and delegates were present from most, if not all, of the Pan American Republics.

The technical papers and discussions were grouped under six sections: (1) Human populations and productive capacity of the land, (2) Renewable resources and international relations, (3) Land use and the social sciences, (4) The dynamics of renewable resources, (5) Education in conservation dynamics, and (6) making conservation effective.

The following papers were deemed to be the most interesting to members of AGU: Conservation and multiple use of water in Mexico, by Adolfo Orive Alba, Secretary of Hydraulic Resources, Mexico, D. F.; Problems in the utilization of water resources in irrigation, by Rodolfo E. Ballester, Professor of Applied Hydraulics, University of Buenos Aires, Buenos Aires, Argentina; Legal and economic problems associated with excessive withdrawal from ground-water sources, by William E. Warne, Assistant Secretary, U. S. Department of the Interior; Geologic implications of watershed management, by Reed W. Bailey, U. S. Forest Service, Ogden, Utah; Soil conservation, by H. H. Bennett, Chief, Soil Conservation Service; Water, by Royal W. Davenport, Hydraulic Engineer, U. S. Geological Survey; Water control and resource development, by James P. Pope, Director, Tennessee Valley Authority; Water limits to human activity, by Gilbert F. White, President, Haverford College, Haverford, Pennsylvania; and Watershed vegetation and the hydrologic cycle, by H. G. Wilm, Conservationist, U. S. Forest Service, New Orleans, Louisiana.

It is anticipated that these papers will be published in full in the proceedings of this Conference.

Seventh Pacific Science Congress held in New Zealand

Charles F. Richter of California Institute of Technology, and Francis P. Shepard of Scripps Institution of Oceanography were the representatives of AGU at the Seventh Pacific Science Congress held in New Zealand in February, 1949. Among the other members of AGU in attendance from the United States were Harold J. Coolidge, K. O. Emery, R. H. Finch, Irving Gringorten, Beno Gutenberg, Robert C. Miller, W. A. Mordy, Robert C. Murphy, B. Perkins, Jr., Clarence S. Ross, Robert H. Simpson, A. F. Spilhaus, Thomas G. Thompson, and Donald E. White.

Robert E. Horton hydrologic papers filed with National Archives

The unpublished notes and documents left to the American Geophysical Union by the late Robert E. Horton have now been deposited with the Archivist of the United States for safe keeping and for access by research students. There are 94 boxes of this material, labelled "National Research Council, American Geophysical Union, Horton Papers." Those contemplating use of these documents should consult a file of the subjects in the office of AGU or in the National Archives.

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